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The Heat Budget of the Arctic Basin

and Its Relation to Climate
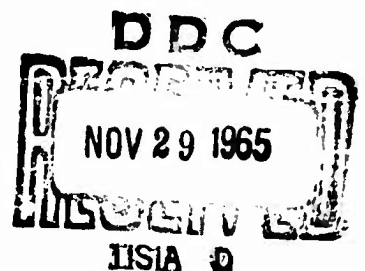
J. O. Fletcher

October 1965

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The Heat Budget of the Arctic Basin
and Its Relation to Climate

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A REPORT PREPARED FOR

UNITED STATES AIR FORCE PROJECT RAND

PREFACE

The study presented here grew out of the author's long interest
in the Arctic and RAND's concern with the possibilities of modifying
the earth's climate, deliberately or inadvertently. It is a quantita-
tive study of the heat budget of the Arctic; the author has sought out,
evaluated, and combined the available data to produce a useful though
rough description of the input, transformation, and output of thermal
energy in the Arctic regions. This quantitative description, together
with the author's long study of, and personal experience in, the Arctic,
has produced new insights into the mechanism of the Arctic climate.

This description of the Arctic heat budget and its relationship
to the atmospheric circulation of the hemisphere provides a spring-
board for imminent numerical experiments that could explain some of
the climatic variations of the past and provide clues as to how the
climate of the future might be foreseen and influenced. The author
concludes by pointing to the specific measurements most urgently
needed in future Arctic investigation.

The work is part of a continuing investigation by RAND in the
area of cloud physics and weather control. It was conducted under
U.S. Air Force Contract AF 49(638)-700.

AUTHOR'S FOREWORD

One purpose of this study is to relate what is known of the Arctic heat budget to what is known of the dynamics of climate.

The study has grown out of four related themes that have long intrigued the writer. Their explicit statement here lends perspective to what follows.

I.

One theme is that men may be able to influence climate on a large scale, and the need to understand how this might be done is rapidly becoming more acute. Climatic variation has always vitally influenced human activity. Year-to-year variations cause good and bad crops over vast regions. Short-term secular trends create grave economic and social problems, as during the drought years of the 1930's. Longer-term trends have often forced drastic changes in human activity, such as the extinction of the flourishing Greenland colonies in the 14th century and the forced migrations from Scandinavia during the "Little Ice Age" of the 17th century.

Yet there is evidence that the changes during the past 3000 years have been mild compared to those in earlier times. It was only some 15,000 years ago that the last great ice sheets began their retreat. Only 5000 years ago, the climate of the Sahara was much like that of present-day Central Europe. Our written history does not record all dislocations of human activity that accompanied these changes, but thick alluvial mud deposits overlying Chaldean cities suggest that the changes were not always gradual.

As the population of the earth will double during the next few decades, both the impact of human activity on climatic balance and the social consequences of climatic change will certainly increase. At

the same time, increasing pressure for the efficient use of natural
resources, including climatic resources, is bound to expand the scale
of man's deliberate intervention in climatic processes. Viewed in
this light, large-scale climatic intervention, either deliberate or
inadvertent, seems probable during our lifetime. Purposeful inter-
vention aimed at better utilization of regional resources is already
in progress (for example, cloud seeding to increase the flow of the
Colorado River). Climatic forecasts and attempts to "stabilize" cli-
mate by countering undesirable secular trends seem just around the
corner. The question is: Can we unravel the complicated interaction
of cause and effect soon enough to avoid costly mistakes?

II.

A second theme is that the atmosphere and the oceans are parts
of a single thermodynamic system in which a change in any part may
influence every other part. Unequal heating and cooling over the
earth drives the circulation of the atmosphere and oceans. If the
heat sources and sinks were removed, the kinetic and available poten-
tial energy of the atmosphere would exhaust itself in a few weeks,
and the system would come to rest. The total effect of altering an
individual heat source or sink may be subtle or insignificant in dis-
tant regions, but -- as will be discussed in Section VI -- it may
not be. The rapid development of numerical models of atmospheric
circulation during the last decade offers very promising tools for
investigating these interactions. In fact, it is the growing belief
of the author that such experiments are already as much limited by
our imperfect knowledge of how the atmosphere is heated and cooled
as by the imperfections of the models.

These reflections emphasize two consequences. One is that
regional alterations of climate should and can be considered in terms
of their effects on the total system. The other is that the problem
is a matter of international rather than national concern.

III.

A third theme is the possibility that the pattern of climate we know is only one of the quasi-stable circulation patterns for the planetary system and that other patterns may be possible. C. E. P. Brooks[*] was first to propose that an ice-free Arctic Ocean corresponds to such a quasi-stable thermal state and that the associated climatic pattern is more "normal" than the one we know, in which a thin film of ice (3 or 4 meters) gives a continental thermal character to the whole vast area separating North America and Eurasia. In recent years Ewing and Donn have elaborated on this theory with a different mechanism to explain the appearance and disappearance of the pack ice.[**]

These discussions have tended to be inconclusive because, lacking a quantitative basis, they have relied on qualitative discussions of dynamic processes that are poorly understood. No one could say with confidence whether the Arctic Ocean would remain ice-free if the pack ice were once removed, or what the response of atmospheric or oceanic circulation would be to such a condition. In fact, the heat budget of the Arctic under the present climatic regime is very poorly understood, and its interaction with atmospheric circulation is even more obscure. To make progress in unraveling these interactions, the central questions are:

1. What is the present heat budget of the earth--atmosphere system in the Arctic?

2. How does this heat budget interact with the present climatic pattern at other latitudes?

3. What is the range and type of natural variation in the Arctic heat budget, and what relationship does this have to the world's climatic anomalies?

4. What would be the heat budget if the ice cover were removed?

5. What would be the new pattern of climate?

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[*] Brooks, C. E. P., Climate Through the Ages, McGraw-Hill Book Company, New York, 1949.

[**] Ewing, M., and W. L. Donn, "A Theory of Ice Ages," Science, Vol. 123, 15 June 1956, pp. 1061--1066; Vol. 127, 16 May 1958, pp. 1159--1162.

These are the central questions, but they suggest others of great practical significance, for after all, if the world's climatic pattern is sensitive to the heat budget of the Arctic, then we should ask:

6. Can observation of heat-budget variations make it possible to anticipate seasonal weather patterns?

7. To what extent could the Arctic heat budget be influenced artificially? What are the prospects for influencing climate in this way?

IV.

The fourth theme is the need for a coherent and purposeful effort to answer these questions. None can be answered with confidence now, but there are direct routes to better understanding that are not being pursued. Progress will be proportional to purposeful effort.

Past programs of Arctic research have too often lacked a coherent guiding theme and have neglected crucial observations.[*] Support for polar research seems often to be governed by precedent and institutional inertia rather than by the need to solve identifiable problems. With these criticisms in mind, the following pattern governs each of the sections of this work: What do we know now? What are the major uncertainties? What more do we need to know? How can we find out?

An objective is to state what should be done, why it should be done and how it could be done, in a way that will capture the interest of physical scientists, be comprehensible to the layman, and be useful to those who guide our national research efforts. My hope is that a coherent overview will allow specific opportunities to be seen in perspective and that perspective will lend purpose and urgency.

Joseph O. Fletcher

Santa Monica, California
September 1965

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[*]Examples are discussed in Section X.

ACKNOWLEDGMENTS

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I. INTRODUCTION AND SUMMARY

There is much empirical evidence relating climatic change to solar variations, but it is not at all clear by what processes such variations could cause the enormous climatic changes associated with the ice ages of the Quaternary Period. After all, variations in solar heating, whether induced by the sun or by dust clouds, must have occurred many times during the last 500 million years. Why should they cause glacial periods only during the last million years?

Evidently there is something special about earth processes during the Quaternary Period. Various authors have referred to "echoing conditions" on the earth or to a "glacial readiness." What this "glacial readiness" consists of has never been established.

It has long been suspected that the ice-covered Arctic Ocean plays some role in this "glacial readiness." On the average, the pack ice is only about 4 meters thick and about 3 million meters across. Removal of the pack ice would change summer albedo from 60 or 80 per cent to 10 or 15 per cent and winter temperatures from -30° to above freezing. Such a drastic difference in surface--atmosphere boundary conditions over such a vast area calls attention to the possible effects on atmospheric circulation. Moreover, abnormal climatic conditions in Europe have long been observed to be associated with abnormal extent of pack ice.

A number of theories for explaining the Quaternary Period's ice ages have made use of these factors. C. E. P. Brooks (1949) in studying the ice ages tried to calculate the critical horizontal extent of pack ice, that is, the size beyond which its own cooling effect would insure its continued growth. Such a condition might represent the critical state between two quasi-stable climatic patterns corresponding to ice-in and ice-out conditions. In 1956 and 1958 Ewing and Donn elaborated

on this theme by advancing an explanation for the formation and dis-
appearance of the pack ice. In their theory the restricted passages
between the Arctic and Atlantic oceans are the critical factors.
Melting glaciers and rising sea level increase oceanic heat exchange
and eventually dissipate the pack ice. Accumulation of continental
glaciers and falling sea levels pinch off oceanic exchange and cause
the pack ice to re-form, thus cutting off the source of moisture for
glacier accumulation.

Now there are two main weaknesses common to these and other
theories for explaining the influence of the pack ice on climate.
One is that without a full quantitative description of the Arctic
heat budget, they tend to be qualitative arguments, difficult to eval-
uate. For example, it is hard to say how plausible Ewing and Donn's
theory is without evaluating quantitatively the relative importance
of oceanic heat transport and the other processes by which heat enters
and leaves the Arctic Ocean. Without detailed knowledge of the heat
budget such comparisons are not possible.

The other weakness is that these theories do not explicitly take
into account changes in atmospheric circulation. There was not any
good way to do so, for few meteorologists would care to gamble on what
the atmospheric circulation pattern would be with an ice-free Arctic
Ocean. Yet atmospheric circulation is the main mechanism by which
heat is transported from one region of the earth to another and cir-
culation patterns do seem to be sensitive even to the variations in
surface heating patterns that occur with our present climate. After
all, the circulation of the atmosphere is forced by the pattern of
heating and cooling. Diabatic heating of the atmosphere produces
available potential energy which is transformed into kinetic energy
by vertical motion, thus maintaining the motion of the atmosphere
against frictional dissipation. Horizontal variations in diabatic
heating create horizontal temperature gradients, which influence the
patterns by which potential energy is transformed into kinetic energy.
In these ways variations in diabatic heating are associated with
variations in general circulation and variations in atmospheric heat
transport.

These interactions have been too poorly understood to be taken
explicitly into account. The usual discussion points out that the
high albedo of the pack ice decreases solar income, permitting greater
atmospheric cooling over the ice and intensifying meridional gradients
and zonal circulation. This reasoning is simple, straightforward, and
logical, but the discussion that follows explains why it is probably
opposite to reality. It sounds logical as a qualitative argument but
when we try to describe it quantitatively we find grave fallacies.

Note to the Reader. The discussion immediately below summarizes
the heat budget in the Arctic that is described in more detail in Sec-
tions II--V and relates this heat budget to certain characteristics of
atmospheric circulation (Section VI). At the same time it carries
along parallel estimates of how each factor would behave if the Arctic
Ocean were free of pack ice. The reader can get a good view of the
form and content of the detailed analysis by following the summary and
studying only those graphs referred to. For some readers this amount
of detail should be enough. Those who are interested in climatic studies
but not in the details of the Arctic heat budget are advised to study
this summary and then turn directly to Section VI. Those who want
to evaluate how these estimates of the heat-budget components are
derived should read Sections II--IV.

HOW HEAT COMES TO AND LEAVES THE CENTRAL ARCTIC

Throughout the year, heat is lost to space by longwave radiation
from the atmosphere and from the earth's surface. This heat loss
cools the atmosphere and surface to temperatures at which the heat
loss is balanced by heat gain from other sources.

During the Arctic's dark season this heat loss is about 10
kcal/cm^2 mo and is balanced by heat from below the surface (about 10
per cent) and from the atmosphere (about 90 per cent). The horizon-
tal thermal gradients so created in the atmosphere play an important
role in forcing the circulation of the atmosphere. Thus in winter
the high rate of atmospheric cooling in high latitudes produces a
pronounced thermal zonality.

In summer the loss of heat to space rises to about 14 kcal/cm^2 mo, but continuous solar radiation compensates for the relatively low elevation of the sun, providing solar radiation at the surface in May, June, and July even greater than at the equator (and much greater than at mid latitudes).

THE NET RADIATION GAINED OR LOST
AT THE SURFACE (RADIATION BALANCE)

Figure 13 (p. 51) shows the behavior of the radiation regime at the surface in the central Arctic during the year. Only about 25 per cent of the solar radiation at the surface is absorbed. This almost equals the net annual loss of longwave radiation at the surface, giving a slightly negative annual radiation balance. The sensitive role played by surface albedo can be seen by noting that a change of only 10 per cent in mean summer albedo would supply enough heat to melt almost 1 meter of ice (about 25 per cent of the whole).

In addition to surface albedo, cloud conditions exercise a powerful influence on the radiation regime. For example, the net loss of longwave radiation at the surface -- shown in Fig. 13 (p. 51) as about 2 kcal/cm^2 mo -- is the difference between outgoing and incoming longwave radiation at the surface. These fluxes are each of the order of 10 kcal/cm^2 mo in winter and 20 kcal/cm^2 mo in summer and are greatly influenced by cloud conditions (Fig. 4, p. 31). Heat income from solar radiation is also sensitive to cloud conditions. It can be seen from Fig. 13 (p. 51) that cloudiness reduces the solar radiation intensity at the surface by about a third. However, by transforming direct solar radiation to diffuse radiation, the fraction absorbed by the high-albedo surface is increased (Fig. 15, p. 55). This sensitivity of both longwave- and solar-radiation components to cloud and albedo conditions suggests ways of observing (and possibly influencing) year-to-year variations in the heat budget.

THE RADIATION BUDGET AT THE SURFACE FOR AN ICE-FREE ARCTIC

If the pack ice were removed, the radiation regime would be drastically different, as can be seen from the estimates of Fig. 16 (p. 57). The annual longwave loss at the surface would be more than doubled, but

the solar radiation absorbed at the surface would be more than tripled, with the annual radiation balance being about 20 kcal/cm^2 greater than at present. The total solar heat absorbed each year at the surface would be equivalent to about twice the heat required for melting the entire present ice pack. To interpret the significance of these numbers, we must examine the other (nonradiative) components of the heat budget.

THE HEAT BUDGET AT THE SURFACE

The radiative components tend to dominate the Arctic heat regime because, relatively, they are so large. At the surface, the sum of all the positive and negative radiative components (the radiation balance) must equal the sum of the heat flows by nonradiative processes: from below (conduction) and from the atmosphere (evaporation and turbulent heat exchange).

These quantities are difficult to measure, and their mean values for the Arctic are relatively uncertain. Data from several sources are discussed, with the conclusion that the results from the Soviet drifting-station observations, as reported by Doronin, are most likely. Heat flow to the surface from below is known with greater reliability than evaporation and sensible-heat exchange. This quantity is taken from the model described by Untersteiner.

Having evaluated various sources of heat-budget data and arrived at estimates of each component separately, the compatibility of the results can be assessed by "balancing the budget" at the surface and noting the resulting discrepancy. This is shown by Fig. 22 (p. 79), which reveals remarkably good consistency among the components. One can conclude that, contrary to the generally accepted notion that not enough is known about the Arctic heat regime to "balance the budget," we can now describe the heat budget at the surface with a degree of uncertainty that is comparable in magnitude to its year-to-year variations.

THE HEAT BUDGET AT THE SURFACE FOR AN ICE-FREE ARCTIC

If the pack ice were removed, the resulting heat budget would be entirely different, as can be seen by comparing Fig. 13 (p. 51) with Fig. 16 (p. 57). This comparison illustrates the immense role played by

the pack ice in suppressing the heat exchange between the surface and deeper layers. In mid winter an ice-free sea would supply heat to the surface at a rate of about 8 kcal/cm^2 mo, about half of which would compensate for the longwave-radiation loss at the surface, and half of which would be lost to the atmosphere by evaporation and turbulent exchange (Fig. 26, p. 93). By contrast, under present pack-ice conditions in mid winter, the surface receives only about 1.5 kcal/cm^2 mo from below, while losing about 2.5 kcal/cm^2 mo in longwave radiation. The surface is cooled until equilibrium is reached. At these temperatures the cold surface drains about 1 kcal/cm^2 mo from the atmosphere, causing the deep surface inversions so characteristic of the Arctic winter.

In summary for an open sea, the annual pattern of evaporation and turbulent heat exchange would be almost opposite to present conditions. Over pack ice, heat flows from the atmosphere to the surface during winter; it flows from the surface to the atmosphere during summer. With an open Arctic sea, however, the winter heat exchange would be about four times as great and in the opposite direction, from the surface to the atmosphere. The summer heat exchange would be somewhat less than at present and in the opposite direction.

Under present conditions in summer about 10 kcal/cm^2 flows down through the surface to deeper layers, mostly for warming the ice mass and melting ice, almost none to warming the ocean. But an ice-free ocean would receive about 40 kcal/cm^2. This heat would be stored in the ocean during summer and would be released during winter to balance the heat losses at the surface. Thus, according to Fig. 26 (p. 93), the heat budget for an ice-free Arctic sea would be slightly positive; that is to say, the pack ice, once removed, might not return.

Actually, the estimates shown in Fig. 26 (p. 93) are quite uncertain, inasmuch as they depend in large part on how atmospheric circulation would adjust to these conditions. As we gain more insight into this factor, we can reduce these uncertainties.

THE PERMANENCE OF THE PACK ICE

From known characteristics of the heat budget of the surface, it is of interest to estimate the effects of abnormal values of summer air

temperature or summer albedo on the thickness and extent of the Arctic pack ice, which at present has a mean thickness of about 4 m.

The most systematic calculations of this sort have been carried out by Soviet investigators (Section V) who report that a 10 per cent decrease of summer albedo (from 70 per cent to 60 per cent) would destroy the pack ice in eight to ten years, whereas a decrease from 70 to 50 per cent would do so in two to three years. A summer air temperature anomaly of $+5^{\circ}C$ would destroy the pack ice in a year and a half to two years; $+4^{\circ}C$ would do so in about four years. Even a $+2^{\circ}C$ anomaly would reduce 4 m of ice to 1 1/2 m in one summer and would destroy the whole pack in a few decades.

THE HEAT BUDGET FOR THE EARTH--ATMOSPHERE SYSTEM

Figure 10 (p. 41) shows the estimated heat budget for the earth--atmosphere system in the central Arctic. The heat loss of the system is entirely represented by the longwave-radiation loss to space, which varies from its mean value by only about 20 per cent during the year. In summer, this loss is largely compensated for by heat gain from solar radiation absorbed at the surface and in the atmosphere. The remainder of the heat loss from the system must be supplied either from below the surface or from the atmosphere.

The most uncertain component in Fig. 10 (p. 41) is the solar radiation absorbed in the Arctic atmosphere. This important component has been studied relatively little by Soviet Arctic investigators and almost not at all by United States Arctic investigators. The values shown are very much larger (100 per cent) than is normally assumed for the earth as a whole. They are taken from fragmentary Soviet sources and are theoretically possible.

These large values are associated with the prevalence over the pack ice in summer of extensive low stratus, whose high transmissivity inter-acts with the high-albedo surface below to produce abnormal absorption in the cloud layer. An investigation of this component by field measure-ment is gravely needed.

THE EARTH--ATMOSPHERE SYSTEM FOR AN ICE-FREE ARCTIC

Figure 17 (p. 59) is the ice-free counterpart of Fig. 10 (p. 41). At first glance they look remarkably similar, but it should be noted that the relative magnitudes of the solar radiation absorbed at the surface and in the atmosphere are reversed, surface absorption being 3 times as large in the ice-free case.

The sum of all the radiation components, which must match the heat supplied to the system by other processes, is rather similar for the two conditions. In winter, the radiative loss is somewhat larger for the ice-free case, but from Fig. 26 (p. 93) we see that this is more than balanced by greater heat flow from the ocean, so the requirement for atmospheric heat would be less than at present. In summer, the greater surplus of radiative heat at the surface for an ice-free Arctic would be more than balanced by the greater heat storage in the ocean (Fig. 26, p. 93), so the summer drain of atmospheric heat would be greater than at present.

Thus, while the net demand for heat from nonradiative sources during the course of the year is rather similar for the two cases, the heat flow from below the surface is drastically different and therefore the heat drain on the atmosphere varies in an inverse way.

ANNUAL INTERACTION OF ARCTIC HEAT BUDGET
AND ATMOSPHERIC CIRCULATION

The requirement for heat from the atmosphere during the course of the year, derived in the above manner, is shown in Fig. 27 (p. 103). The meager data available from direct measurements of atmospheric heat advection are also shown and are in good agreement.

The lower part of the figure shows the observed variation in several related indices of atmospheric circulation intensity during the year. During mid winter, when demand for atmospheric heat is greatest in high latitudes, these indices all are at their maximum, and in summer, when demand is least, they are at their minimum.

In a similar way, the demand on the atmosphere for heat can be estimated for an open Arctic Ocean (Fig. 27, p. 103). The curve in the figure is remarkable for its relative constancy during the year. In

January, the demand on the atmosphere would be only half as great as under present conditions; in summer, the demand would be about the same as in January, whereas under present conditions it approaches zero in June.

The question of how atmospheric circulation would react to this different pattern of atmospheric cooling calls attention to the link between the Arctic heat budget and the climate at lower latitudes. Actually we know far too little about the causes of atmospheric circulation patterns to answer this question with any confidence. However, three different approaches are of interest:

1. Qualitative reasoning about synoptic processes and cyclone tracks;

2. Examination of observed changes in atmospheric circulation and ice extent during recent centuries;

3. Experiments with numerical models of atmospheric circulation.

Qualitative Reasoning

The role of the pack ice in climatic change has been discussed by many authors. The usual discussion points out that the high albedo of the pack ice, decreases solar income, increases regional heat deficit, cools the atmosphere, and intensifies meridional temperature gradients and zonal circulation. Usually no distinction is made between summer and winter, but inasmuch as solar heating is the issue, the effect should be most evident in summer.

In fact, the most striking feature of Fig. 27 (p. 103) is the relatively high rate of atmospheric cooling in summer for the ice-out condition. Contrary to the usual argument, we should actually expect in summer greater meridional temperature gradients and more vigorous atmospheric circulation for the ice-out condition than for the ice-in condition. It is true that solar income is very high for the earth--atmosphere system, but most of this income is being stored as oceanic heat -- to be released to the atmosphere during winter.

During winter an open Arctic Ocean furnishes much of the heat loss of the earth--atmosphere system, and atmospheric cooling would be only about half as great as with an ice cover. This does imply some decrease in vigor of winter circulation but probably not much.

We must remember that the temperature _difference_ betwe··n une air entering the region and the air leaving the region will be u.::ch smaller. If it is half as great, then, the same vigor of circulation that we have now would supply only half as much heat.

Thus we may conclude that for the year as a whole an ice-free Arctic Ocean would correspond to more vigorous atmospheric circulation and that this effect would be most pronounced in summer.

Considering the annual sums of atmospheric cooling, we may note that the difference between ice in and ice out is only 15 $kcal/cm^2$, a reduction of about 20 per cent from the ice-in condition. Such a small change in the annual sum is not likely to be very significant to the heat balance of the hemisphere or to counterbalance the tendencies to more vigorous circulation noted above. What is striking about the ice-out condition is the relative constancy during the year, implying a monsoon type of circulation in high latitudes during the whole year.

Thus the atmospheric circulation associated with an ice-free Arctic would be highly conducive to glacier formation, with cool, moist conditions and vigorous zonal circulation at high latitudes during the whole year. The increased advection of heat to the Arctic in summer should give cooler summers at mid latitudes, thus reducing the ablation of snow. As more land area became covered with snow in summer the higher albedo would cause even further cooling.

The main hemispheric centers of action should not be greatly affected, but in winter the Arctic Ocean would probably be an area of substantial cyclogenesis rather than anticyclogenesis as at present. The ridge between Alaska and Greenland would be replaced by a trough. The two avenues by which cyclones enter the Arctic Basin would still be the Norwegian Sea and the Bering Sea, but the relative importance of the latter would increase, with more cyclones of the northwest Pacific being deflected north into the Arctic rather than south across North America as at present. Thus the central United States would have dry winters. Over the Arctic Basin the moisture-laden cyclones would be further fed by surface heat and moisture (about 30 cm of water), resulting in great instability and precipitation, in marked contrast to present conditions. Very heavy snowfall would be expected

in northern Canada, where the monsoonal winter circulation and the high terrain would provide exceptionally favorable conditions for glacial accumulation. The Pleistocene ice sheets could have originated in this way.

In summer, the much greater cooling of the Arctic atmosphere (Fig. 27, p.103) should be reflected in more intense meridional (and zonal) circulation, probably by a tendency for North Atlantic cyclones to take more northward paths, giving Europe dry summers.

Thus an ice-free Arctic would probably be associated with a monsoonal circulation both summer and winter, with greater aridity for the United States and Europe and heavy snowfall in northern Canada and northern Scandinavia.

In the absence of a suitable model, only crude estimates can be made of climatic conditions associated with an ice-free Arctic. According to Soviet estimates, mid latitudes would be 10--$13^{o}C$ warmer than at present and low latitudes $5^{o}C$ cooler in winter. In the central Arctic, air temperatures would be about $+5^{o}C$ in summer, whereas water temperatures in the coldest month would be near $+5^{o}C$. In the peripheral zone, mean surface temperatures would be around $-10^{o}C$ in winter and $+10$--$15^{o}C$ in summer, with shore ice building out during winter and disappearing in May.

Observed Secular Changes

Observational evidence indicates that the intensity of atmospheric circulation has undergone substantial variations, even during the last century and that these variations seem to coincide with variations in ice extent in Arctic seas. Some of these data, as presented by Lamb, are shown in Figs. 38 (p.123) and 39 (p.124).

According to these data, maximum intensity of atmospheric circulation occurs when the area of pack ice is least. It is of interest to ask, "Does less ice operate to increase or decrease atmospheric circulation?" According to the usual qualitative argument, less ice operates to decrease and is thus a stabilizing influence and not a causal factor. According to our estimates less ice operates to increase circulation and is thus a causal factor.

Climatic Experiments

The foregoing evidence suggests a number of possible relationships between Arctic heat budget and atmospheric circulation, but such qualitative reasoning and statistical correlation are very unreliable methods for distinguishing between cause and effect. The interaction of the atmosphere with the many dynamic and thermal factors that influence its motion is so complicated that qualitative methods can do little more than suggest possibilities. Fortunately, numerical methods, newly developed for integrating the primitive equations of atmospheric motion, are now beginning to provide models with which quantitative experiments can be conducted. For example, with a model that simulates the behavior of the real atmosphere one can experimentally investigate the effects of changing a particular boundary condition simply by changing that condition and observing the reaction of the atmospheric circulation. Section IX suggests a series of such experiments aimed at unraveling the cause-and-effect relationships. With the present crude models that are available and our relatively poor knowledge of the heat budget and heat-exchange processes, only the most extreme anomalies can be usefully investigated. These include such extremes as pack ice present or absent and continental glaciers present or absent. To examine combinations of these extremes for both summer and winter would require eight experiments (including the "control" experiment representing present conditions). The results should give new insights on the reaction of atmospheric circulation to each of these conditions, pointing the way to the interpretation of observational data on past climates and to the design of more sophisticated experiments.

OCEANIC CIRCULATION

Compared with the other components of the Arctic heat budget, the heat transport by oceanic circulation into the Arctic Ocean is relatively small, being comparable to about 5 per cent change in mean surface albedo (Fig. 42, p.133). However, the importance of oceanic circulation could be very much greater than these numbers suggest, for the pattern of oceanic heating exercises strong leverage on the extent

of pack ice and on cyclone tracks, both factors accounting for large magnitudes of heat. As shown in Fig. 45 (p. 138), most of the heat entering the Arctic by oceanic advection is dissipated upward near the margins of the pack ice. Thus, anomalies in oceanic heat advection can greatly affect ice extent in this region.

The pattern of air--sea heat flow and the paths of cyclones are affected both by the extent of pack ice and by anomalies of water temperature (Fig. 36, p. 120). Conversely, the extent of pack ice is sensitive to the temperature of air masses. The complicated inter-action of these three factors, ice extent, oceanic advection, and atmospheric advection requires much further investigation.

Present models of atmospheric circulation are probably too crude for investigating anomalies in these factors of a magnitude observed in nature, but expected improvements should make such experiments feasible in a few years. Meanwhile, more complete observation of how nature behaves would improve the basis for such investigation. In particular, measurement of the transport of water, ice, heat, and salt between Greenland and Spitsbergen during the course of the year would greatly improve understanding of oceanic exchange.

On the other hand, the experiments outlined above involving the more extreme conditions should give new insight into oceanic cir-culation associated with past climates, for the driving force is wind stress on the ocean surface, and the mean pattern of oceanic circula-tion is closely related to the mean atmospheric pressure pattern. The mean pressure patterns obtained experimentally should indicate features of oceanic circulation associated with past climates.

CLIMATE PREDICTION AND CONTROL

Since atmospheric circulation is forced by the pattern of heating and cooling over the earth, it should be possible to predict certain characteristics of future atmospheric patterns from knowledge of future heating patterns. In an empirical way, man has always done this. For example, winter is different from summer because the solar heating pattern is different. Many characteristics of the January climate are known in advance, but we are still not able to predict

how one January will differ from another, because we do not yet under-
stand what causes one January to be different from another. Improve-
ment in our ability to predict climate involves, on the one hand,
better understanding of how the atmosphere interacts with certain
heating patterns, and on the other hand, better understanding of why
these patterns exist and how they will be different in the future.

Until now, research in climatic theory has employed mainly
synoptic and statistical methods and has been largely descriptive.
These techniques are now about to be supplemented by powerful experi-
mental tools, but the utilization of these tools requires a better
quantitative knowledge of heat budget and of surface--atmosphere
interaction.

The major objective of this study is to assess quantitatively the
various components and processes of the Arctic heat budget to (a)
encourage climatic experiments in the laboratory; (b) provide quanti-
tative inputs to such experiments; and (c) identify and assess the
main uncertainties in our knowledge of the Arctic heat budget and
suggest how to reduce them.

Our ability to predict climatic changes will grow as we find out
how the atmosphere responds to various heating patterns. To the
extent that these heating patterns could be influenced artificially,
purposeful climate control would be feasible. Section VIII discusses
possible ways of influencing each of the components of the Arctic
heat budget. The conclusion is that even present engineering capa-
bilities could influence the heat budget enough to be climatically
significant; it might even be possible to remove the pack ice. Thus
our gross capability to influence is much greater than our under-
standing of how to do so purposefully.

NEEDED ARCTIC RESEARCH

Section X outlines a number of ways to reduce quantitative
uncertainties about the Arctic heat budget. The greatest single
uncertainty is the influence of cloudiness on the absorption of solar
radiation in the atmosphere. This is probably an important uncertainty
in all regions but is more acute in the Arctic because of continuous

summer sunlight and because probably nowhere else on earth does such
a simple and uniform cloud condition persist over such a wide extent
for so much of the time.

Other important field measurements to be emphasized include time-
and-space variations in surface albedo during summer and variations in
cloud conditions during the whole year.

In some cases important data sources already exist but have not
been utilized. Examples include almost two decades of cloud data
from weather reconnaissance flights over the Arctic and a rather good
radiosonde network, from which computations of atmospheric heat advec-
tion could be made.

II. THE RADIATION CLIMATE OVER PACK ICE

INTRODUCTION

The heat exchange between earth and space is maintained by two great streams of radiant energy, incoming solar radiation and outgoing longwave radiation from the earth's surface and atmosphere. These radiant fluxes are usually much larger than other terms in the energy budget, such as evaporation, sensible heat exchange, and advection. Their behavior during the year thus determines the main features of the heat budget, which includes all the components of heat flow.

In this section we will describe the behavior of each radiation component during the course of the year. Each component is expressed as the rate at which energy falls upon (or leaves) a unit area of horizontal surface. In most cases it is convenient to use kilo-calories per centimeter squared per month as units.

Since there is no universal convention for designating these components, the following definitions have been adopted for this study.

The Solar-Radiation Components

Above the atmosphere, the intensity of solar radiation is simply the solar constant (about 85 $kcal/cm^2$ mo) times the sine of the solar elevation angle. In passing through the atmosphere some of this energy, q', is absorbed by clouds, water vapor, carbon dioxide, ozone or other atmospheric constituents, some reaches the surface as direct solar radiation, Q, and some reaches the surface as scattered solar radiation, q. The total solar-radiation intensity at the surface is represented by the term $Q + q$.

It is sometimes convenient to consider the amount of solar energy that would reach the surface in a cloudless atmosphere. This term is called total possible solar radiation at the surface and is designated by $(Q + q)_0$. It takes into account solar elevation, scattering from a cloudless atmosphere, and surface reflection.

At the surface a fraction, α, of the incident solar radiation is reflected, and the remaining fraction, $1 - \alpha$, is absorbed. Thus the solar radiation absorbed at the surface is represented by $(Q + q)(1 - \alpha)$. The term α is called the albedo <u>of the surface</u>.

The Longwave-Radiation Components

Although a cloudless atmosphere is relatively transparent to solar radiation that lies mostly in the spectral region 0.3 to 0.5 microns,[1] it is relatively opaque to the longwave radiation originating from bodies radiating at terrestrial temperatures. This radiation (4 to 18 microns) originates from the surface, from clouds, and from the minor atmospheric constituents. Water vapor and carbon dioxide both absorb and radiate strongly in this region,[2] so only about one fourth of the longwave radiation from the surface can pass directly through even a cloudless sky to escape to space. The rest is absorbed and reradiated, some of it back to the surface. With an overcast sky, almost all of the radiation from the surface is absorbed, back radiation from the atmosphere to the surface tends to be high, and the longwave loss to space comes almost entirely from the atmosphere.

The net longwave-radiation at the surface is the difference between the outgoing longwave radiation and the longwave radiation from the atmosphere that is absorbed at the surface. This quantity is designated by I and represents the rate at which the surface is losing heat in its longwave radiative exchange with the atmosphere and space.

The total longwave loss to space at the top of the atmosphere is designated as I_∞. Most of this radiation has originated from the atmospheric constituents, but as noted above, some may have originated at the surface and passed through the atmosphere unimpeded.

THE RADIATION-BALANCE EQUATIONS

The net heating or cooling at the surface due to all the radiation components is commonly called the radiation balance. (It would be more physically meaningful to call this term the radiation imbalance, for it represents the resultant of all the radiant fluxes and must equal the heat reaching or leaving the surface by nonradiative processes.) This quantity is designated F_r and can be represented by the formula:

$$F_r = \text{net solar heating} - \text{net longwave cooling}$$

$$= (Q + q)(1 - \alpha) - I \quad , \tag{1}$$

where:

F_r = radiation balance at the surface,[*]

Q and q = direct and diffuse fluxes of solar radiation at the surface,

α = surface albedo,

I = net longwave-radiation loss at the surface.

The rate of radiant heat loss or gain by the atmosphere is called the radiation balance of the atmosphere; it is designated by B_a and can be expressed by:

$$B_a = \text{net longwave gain from the surface} - \text{total}$$
$$\text{longwave loss to space} + \text{solar radiation}$$
$$\text{absorbed in the atmosphere; i.e.,}$$

$$= I - I_\infty + q' \quad . \tag{2}$$

The rate of radiant heat loss or gain for the earth--atmosphere system can be expressed as the sum of the radiation balance of the surface and the radiation balance of the atmosphere, described above by Eqs. (1) and (2).

$$B_{e/a} = (Q + q)(1 - \alpha) + q' - I_\infty \tag{3}$$

$$= \text{solar absorption at the surface} + \text{solar}$$
$$\text{absorption by the atmosphere} - \text{longwave}$$
$$\text{loss to space.}$$

---

[*]The notation F_r follows the example of Badgley and Untersteiner. In Soviet literature R is usually used, and B_e is sometimes used to represent radiation balance of the surface.

It may be noted that during the dark portion of the year all the
solar components are zero and Eqs. (1), (2), and (3) reduce to:

$$F_r = -I$$
$$B_a = I - I_\infty \qquad\qquad (4)$$
$$B_{e/a} = -I_\infty \quad .$$

SOURCES OF DATA

In the Soviet Arctic, data are available from about five perma-
nent actinometric stations for a period of 10--20 years and from more
than twenty stations for about 10 years. In the American Arctic, total
solar radiation at the surface has been measured at about ten stations
for about 10 years. Longwave components have been measured sporadi-
cally at a few stations during and since the International Geophysical
Year.

On the whole, the solar-radiation components have been much more
extensively measured than the longwave components. The latter are
usually calculated from measured meteorological variables.

Attempts have been made by various investigators to compile grid-
point data for the Arctic to include all the radiation components. In
the Soviet Union the data reported by Marshunova[3] and Gavrilova[4]
are especially comprehensive. Outside of the Soviet Union, the most
comprehensive attempt has been carried out by S. Orvig, E. Vowinckel
and others at McGill University.[5--11] The greatest number of non-
Soviet field measurements on the Arctic heat budget have been made by
scientists at the University of Washington.

After careful consideration, available Soviet data were chosen
as the primary basis for this study. These data, on the whole, seem
more internally consistent, are more complete and are supported by
more field observations, than data from any other source. Where major
discrepancies exist between these data and those reported from McGill
University or from the University of Washington, the conflicting
results are compared and discussed.

The remainder of Section II describes the behavior of each of
these components during the course of a year. The quantitative

estimates are based mainly on data reported by Marshunova[3] and Gavrilova.[4] Their results represent a combination of observed and calculated data.

QUANTITATIVE UNCERTAINTIES

On the basis of Soviet data, Gavrilova estimates that total solar radiation in the Arctic is known to 2--5 per cent for annual values, and 5--10 per cent for monthly values, absorbed solar radiation to 10--15 per cent, net outgoing longwave radiation from the surface to 15--20 per cent, and the radiation balance at the surface to 20--30 per cent. This estimate may be optimistic (for the McGill data differ from Soviet data by a much wider factor), but it gives a reasonable perspective for viewing the data that follow.

TOTAL POSSIBLE SOLAR RADIATION AT THE SURFACE UNDER A CLEAR SKY

The term $Q + q$ in Eq. (1) represents the total direct and diffuse solar radiation at the surface after some reduction by clouds. The maximum possible value corresponds to clear skies and is represented by $(Q + q)_0$.

It is a popular belief that the solar energy falling on a given horizontal surface area is smaller at the poles than at the equator because the sun's elevation is low. Actually, this is not true in summer, when the longer hours of sunlight result in about 20 per cent more energy per day falling on the surface than in temperate latitudes. Figure 1 and Table 1 give values of possible total solar radiation $(Q + q)_0$ according to Gavrilova and Marshunova, respectively.

THE EFFECTS OF CLOUDINESS ON INCOMING SOLAR RADIATION

The prevailing cloud condition in the Arctic Basin during summer is a thin low stratus, whose small density reduces the total short-wave radiation at the surface only moderately, though the direct radiation is scattered and diffuse radiation is thus increased. Figure 2 shows the typical relationship between direct, diffuse, and total solar radiation at the surface. The sudden onset of stratus

Table 1

THE CENTRAL ARCTIC:

POSSIBLE TOTAL RADIATION AT THE SURFACE UNDER A CLEAR SKY, $(Q + q)_0^*$

$(kcal/cm^2)$

North Latitude	Month									
	Feb	Mar	Apr	May	June	July	Aug	Sept	Oct	Nov
90°	...	0.2	9.5	22.0	25.8	25.0	15.6	3.0	...	...
88°	...	0.2	9.5	22.0	25.8	25.0	15.6	3.2	...	...
86°	...	0.4	10.0	22.0	25.8	25.0	15.6	3.4	...	...
84°	...	1.1	10.3	22.0	25.8	25.0	16.0	3.6	...	...
82°	...	1.6	10.6	22.0	25.8	25.0	16.0	4.2	0.2	...
80°	...	2.1	11.0	21.8	25.7	25.0	16.0	5.0	0.5	...
78°	0.2	3.0	11.5	21.3	25.6	24.5	16.2	5.7	0.7	...
76°	0.3	3.6	11.8	20.8	25.1	24.0	16.2	6.5	1.0	...
74°	0.5	4.4	12.4	20.4	24.4	23.3	16.8	7.7	1.6	...

*After Marshunova.
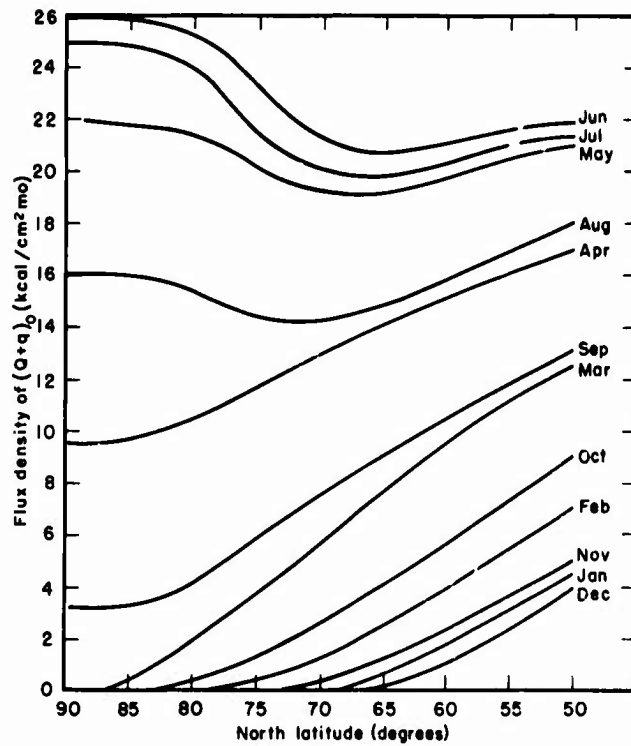
Fig. 1 -- Total solar radiation falling on a horizontal surface
under a clear sky $(Q + q)_0$ (after Budyko).
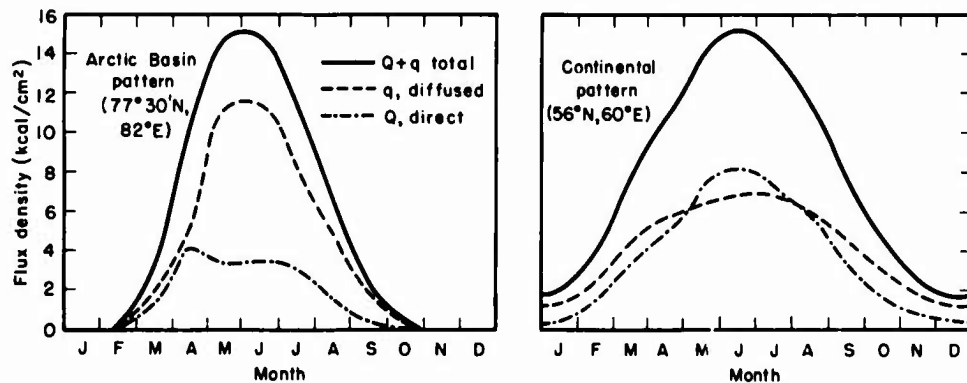
Fig. 2 -- The effects of cloudiness on the amount of solar
radiation reaching the surface (after Gavrilova).

clouds over the pack ice in early April, as air masses from warmer
land areas move over the ice, is marked by a sharp change in the
character of the radiation reaching the surface.

The high surface albedo also increases the intensity of diffuse
solar radiation falling on the surface, for some of the solar radia-
tion reflected upward from the surface is scattered back down on to
the surface. For this reason, the total solar-radiation intensity on
the surface, Q + q, is about 20 per cent greater over pack ice (albedo
70--90 per cent) than over open water (albedo 10--15 per cent).

In the central Arctic, the direct radiation, Q, is only 20--30
per cent of the total, Q + q. The relative intensity of diffuse
radiation, q, is so high that the annual accumulation of diffuse
radiation for the central Arctic is comparable to values for mid
latitudes, in spite of the long Arctic night.

These conditions are generally characteristic of the whole pack-
ice area; they vary little from place to place, as can be seen from
Table 2, which summarizes values of Q + q based on field observations.
Comparison of Table 2, Q + q, with Table 1, $(Q + q)_0$, shows that the
cloudiness does not reduce the solar-radiation intensity at the sur-
face as much as one might expect.

The high intensity of solar radiation partially trapped between
a highly reflecting surface and a highly scattering cloud layer results
in a greater absorption of solar radiation in the atmosphere than is
typical of other regions. This factor will be discussed more fully
later.

THE EFFECTS OF SURFACE ALBEDO ON SOLAR RADIATION

At the high-albedo surface, only a small fraction of the total rad-
iation is absorbed. Figure 3 illustrates this relationship for a rela-
tively cloudy area (Is Fjorden) and a relatively clear area (Isachsen).
These locations may be found on the map at the beginning of Section I.

From sunrise (February--March) until June, only a small fraction
of available solar radiation is absorbed. The abrupt decrease in
albedo in mid summer corresponds to the collapse of the snow cover
and the appearance of dirt or puddles on the ice. By the time this

Table 2

CENTRAL ARCTIC:

TOTAL SOLAR-RADIATION INTENSITY AT THE SURFACE, Q + q[*]

Lati-tude	Longi-tude	Month										Annual Total
		Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	
		Intensity (kcal/cm^2)										
90°N		...	...	8.1	17.6	19.2	13.5	8.3	2.2	...	...	68.9
85°N	0°E	...	0.6	8.5	17.6	19.2	13.5	8.4	2.5	...	...	70.3
85°N	60°E	...	0.6	8.4	17.6	19.2	13.5	8.2	2.5	...	...	70.0
85°N	120°E	...	0.6	8.4	17.6	19.2	13.5	8.4	2.6	...	...	70.3
85°N	180°E	...	0.6	8.6	17.6	19.2	13.5	8.4	2.6	...	...	70.5
85°N	120°W	...	0.6	8.6	17.8	19.2	14.0	8.5	2.6	...	...	71.3
80°N	135°E	...	1.8	10.1	17.5	19.1	13.7	8.6	3.6	0.4	...	74.8
80°N	165°E	...	1.9	10.0	17.7	19.0	14.0	8.6	3.6	0.4	...	75.2
80°N	165°W	...	1.9	9.9	17.7	19.0	14.0	8.6	3.7	0.4	...	75.2
80°N	135°W	...	1.9	9.9	17.6	19.2	14.2	9.0	3.7	0.4	...	75.9
75°N	160°E	0.4	3.6	10.0	16.6	18.5	13.7	9.2	5.2	1.1	...	78.3
75°N	180°E	0.4	3.6	10.0	16.8	18.5	13.5	9.2	5.2	1.1	...	78.3
75°N	160°W	0.4	3.6	10.3	16.8	18.5	13.7	9.2	5.2	1.1	...	78.8
75°N	140°W	0.4	3.6	10.3	16.8	18.7	13.7	9.2	5.3	1.1	...	79.1

[*]After Marshunova.

happens the available solar radiation is rapidly decreasing. Thus the heat budget for the season is very sensitive to the date puddling begins. If this date were advanced by a month, for example, the seasonal radiation income at the surface would be drastically increased. Actually this date does vary from year to year, and the corresponding variations in heat intake must influence ice conditions and possibly weather patterns for the following winter. Systematic observations of albedo over the Arctic Basin combined with systematic observations of the total ice budget, may have forecasting value and may provide insight into possible ways of influencing climate. This will be discussed more fully in later sections.

Table 3 shows change of albedo with season for several Arctic areas. The highest values (80--85 per cent) are typical for dry snow, whereas over open seas the values may be as low as 10 per cent.

Table 4 shows absorbed solar radiation for the central Arctic. It is of interest that both pattern and annual total vary little from place to place, as long as the underlying surface is pack ice.

THE LONGWAVE-RADIATION COMPONENTS

A clear atmosphere absorbs about three fourths of the longwave radiation from the surface. Clouds scatter and absorb this radiation even in the 8μ--10μ atmospheric "window"; consequently, clouds and atmosphere together absorb almost all the ground's longwave radiation. This absorbed energy is reradiated back toward the ground and outward toward space.[*]

The longwave balance in the atmosphere and that at the surface are controlled by the differences in temperature between the surface and the cloud tops or bottoms. Cloud tops cooler than the ground may radiate less energy to space even though the relatively dry atmosphere above the clouds lets more longwave radiation through. Cloud tops warmer than the ground, common in Arctic winters, lose more to space than does the ground under a clear sky.

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[*]Except for the cirrus it can be assumed that clouds radiate as black bodies or almost so. Marshunova reports emissivity of 0.85 at $-26°C$, increasing to 1.0 at $-4°C$. This is a factor that is very poorly known and for which few measurements have been taken.

Table 3

SURFACE ALBEDO, α, AS A FUNCTION OF SEASON[*]

(per cent)

Location	Month									
	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov
Drifting Stations	83	83	81	82	78	64	69	84	85	...
Tikhaya Bay	...	83	83	83	70	40	22	22	47	...
Dickson Island	86	87	85	80	41	20	27	35	82	86
Tiksi Bay	86	86	86	80	35	30	30	30	66	86
Cape Schmidt	80	80	80	80	43	21	20	44	46	80
Kara Sea										
SW	82	80	79	78	54	24	12	10	15	80
NE	82	80	79	78	54	35	35	15	40	...
Lapte Sea	85	82	79	76	55	35	30	25	40	...
E. Siberian Sea	85	82	80	76	70	50	30	25	80	80
Chukchi Sea	82	81	78	70	40	23	20	15	20	70

[*]After Marshunova.
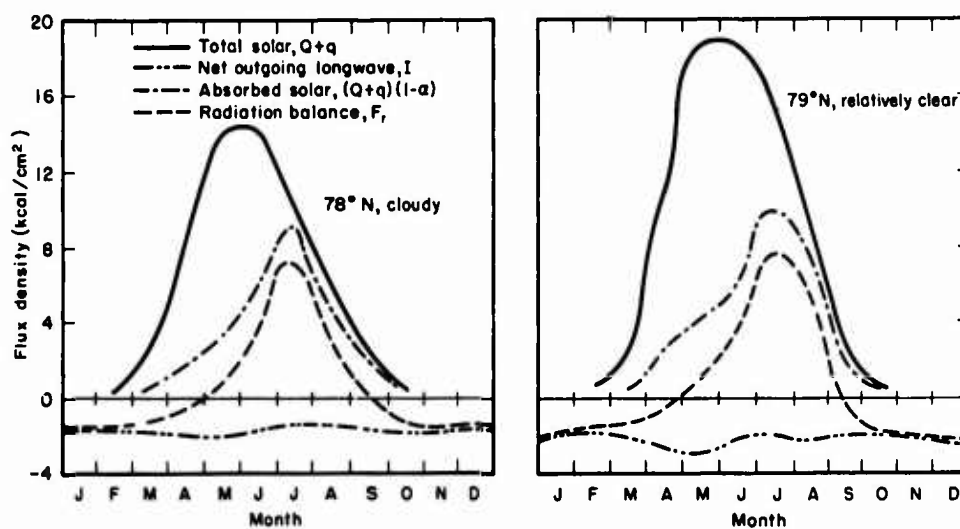
Fig. 3 -- Solar radiation components at
the surface (after Gavrilova).

Table 4

SHORTWAVE RADIATION ABSORBED AT THE SURFACE, $(Q + q)(1 - \alpha)$,

IN THE CENTRAL ARCTIC[*]

$(kcal/cm^2)$

Location	Month									
	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov
90°N	...	...	1.6	3.2	4.2	4.9	2.6	0.4	...	...
85°N	...	0.1	1.6	3.2	4.2	4.9	2.6	0.4	...	...
85°N, 60°E	...	0.1	1.6	3.2	4.2	4.9	2.6	0.4	...	...
85°N, 120°E	...	0.1	1.6	3.2	4.2	4.9	2.6	0.4	...	...
85°N, 180°E	...	0.1	1.6	3.2	4.2	4.9	2.6	0.4	...	...
85°N, 120°W	...	0.1	1.6	3.2	4.2	5.0	2.6	0.4	...	...
80°N, 135°E	...	0.3	1.9	3.1	4.1	4.9	2.7	0.6	...	...
80°N, 165°E	...	0.3	1.9	3.2	4.1	5.0	2.7	0.6	...	...
80°N, 165°W	...	0.3	1.9	3.3	4.2	5.0	2.7	0.6	...	...
80°N, 135°W	...	0.3	1.9	3.2	4.2	5.1	2.8	0.6	...	...
75°N, 160°E	0.1	0.6	1.9	3.0	4.1	5.0	2.9	0.8	0.2	...
75°N, 180°E	0.1	0.6	1.9	3.0	4.1	4.9	2.9	0.8	0.2	...
75°N, 160°W	0.1	0.6	2.0	3.0	4.1	5.0	2.9	0.8	0.2	...
75°N, 140°W	0.1	0.6	2.0	3.0	4.1	5.0	2.9	0.9	0.2	...

[*]After Marshunova.

Reradiation from clouds raises the longwave-radiation balance at the surface, and when cloud bottoms are warmer than the surface, the balance may even be positive. Furthermore, if a temperature inversion near the surface is deep, the longwave-radiation balance at the surface may even be positive without clouds; that is to say, the downward longwave radiation from the warmer air more than compensates for the loss of surface radiation including that in the "window" region.

Thus, in general, the presence of clouds decreases the longwave-radiation loss, I, at the surface and increases the longwave-radiation loss, $I - I_\infty$, of the atmosphere, while the loss to space may be either greater or less depending on relative radiation temperatures of cloud tops compared to the surface.

The calculation of representative values for the longwave components involves knowledge not only of meteorological conditions at the surface but also of the temperature variation and the humidity distribution with height and the extent of low, medium, and high cloudiness. Such calculations have been carried out by several Soviet authors and by Vowinckel and Orvig at McGill University.[5--11] The author believes the results reported by Marshunova to be most reliable, and they are used here. Vowinckel's results are in generally good agreement.

Figure 4 shows the mean values of longwave components in the central Arctic. It was constructed by taking the mean of monthly values reported by Marshunova for drifting stations North Pole 4, 1955--56, North Pole 4, 1956--57, North Pole 5, 1956--57, and North Pole 6, 1956--57.

Figure 4 illustrates the relative stability of the net long-wave-radiation components during the year and the sensitivity of the net longwave radiation to cloud conditions. During January, February, and March there are relatively few clouds, but net outgoing longwave radiation is limited by low surface temperatures. From April to mid summer surface temperatures are rapidly rising but back-radiation from increasing cloud cover and greater atmospheric moisture more than compensate for the greater outgoing radiation at the surface.

The net loss is actually reduced and is minimum at mid summer, when 80--90 per cent low cloud cover is typical.[*]

THE RADIATION BALANCE AT THE SURFACE

The Appendix is a compilation of the components of the radiation balance of the surface for various grid points in the Arctic Basin, according to Marshunova. This compilation is based on calculations from meteorological parameters but has been carefully compared with the aggregate of Soviet radiation measurements. The values given are represented as mean multi-year monthly values, since calculations were made from mean meteorological parameters. It is probably the best representation of the radiation climate of the central Arctic that is available at this time.

The most significant features of the radiation regime can be seen most readily by plotting and comparing some of these data as follows:

Figure 5 shows the components of the radiation balance of the surface for 90°N and Fig. 6 for 75°N, 180°E, a point with ice cover the whole year. Superimposing the two figures shows strikingly that the various components follow almost identical courses, even though the two locations are separated by 15° latitude (900 n mi). This comparison demonstrates that the radiation regime is basically controlled by the character of the surface and by mean cloud conditions rather than by location. The most significant difference between the two figures is the slightly greater net outgoing longwave radiation in mid winter at 75°N, 180°E, reflecting the somewhat greater cloudiness and less extreme inversions of the surface at that location. Either of these figures could be taken as approximately representative of the whole pack-ice region. The differences are within the uncertainty of the data and their year-to-year variations. At both locations the high albedo of the surface causes about 76 per cent of the incident shortwave radiation to be reflected. At both locations the albedo does fall when puddling occurs on the ice, and the highest radiation balance occurs well after the maximum of solar

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[*]To see what this relationship would be like under clear sky conditions, see the 0-km curves in Fig. 8.
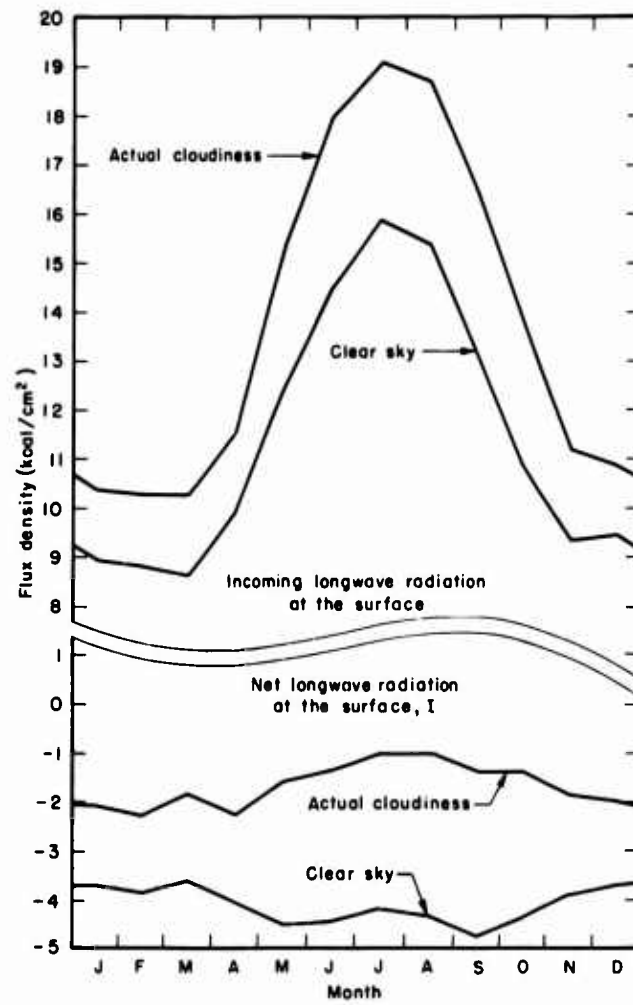
Fig. 4 -- Longwave-radiation components at the surface.

radiation. However, in both cases the decrease in albedo occurs too late in the season to make a big difference in the net absorption.

Superimposing Fig. 6 on Fig. 7 shows the striking effect of substituting a water surface for an ice surface. Figure 7 shows the mean radiation components at the surface for ice-free points in the Barents Sea.[*] This is about the same latitude as Fig. 6 but differs by being more closely observed and being almost free of ice. The greater cloudiness substantially reduces the total shortwave radiation reaching the surface,[**] but the albedo of the water surface is so low that almost all of the solar energy reaching the surface is absorbed. The result is a much higher annual radiation balance, F_r, at the surface, in spite of the high longwave losses from the water surface during winter.

TH RADIATION BALANCE OF THE ATMOSPHERE

The radiation balance for the atmosphere as a whole is represented by Eq. (2):

$$B_a = I - I_\infty + q' .$$

Except in mid summer the shortwave radiation absorbed in the atmosphere, q', is relatively small, and the radiation balance of the atmosphere is basically determined by the fluxes of longwave radiation. The variation of these fluxes with height depends on the temperature and humidity stratification.

This dependence is illustrated in Fig. 8. The values shown were calculated by Marshunova for clear-sky conditions, using mean values of temperature and humidity. The effects of inversions in the first few kilometers are clearly evident.

In general, downward flux decreases with height. In summer, the rate of decrease also decreases with height. In winter, however, the rate of decrease is greatest at 2--4 km. During the late winter,

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[*]After Gavrilova, mean values for Stations 9, 10, 11, Ref. 4, App. 5, p. 175.

[**]According to Budyko,[12] this area, because of great cloudiness, has about the smallest value of total radiation, $Q + q$, on the earth.
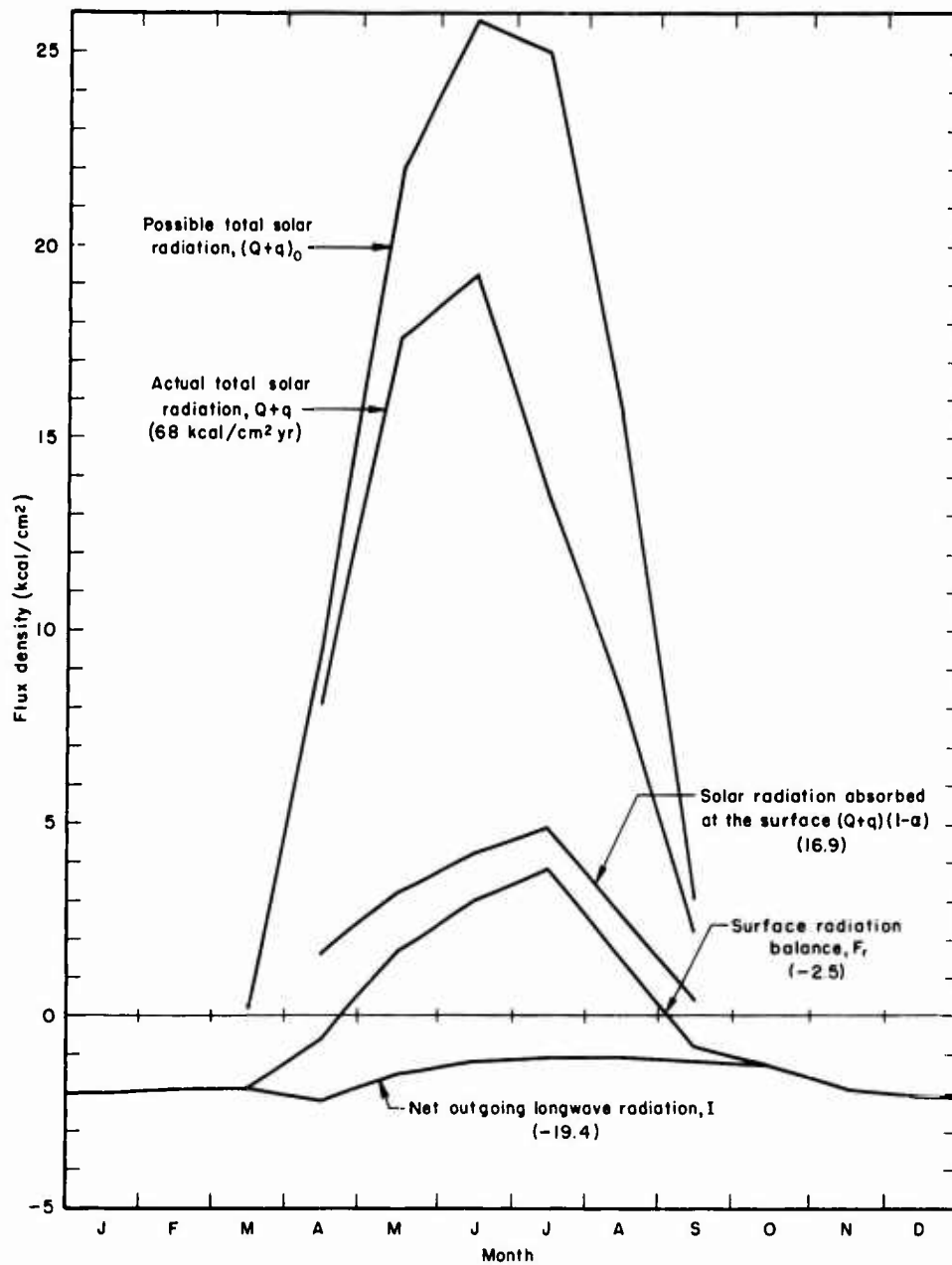
Fig. 5 -- Components of the radiation balance at the
surface at 90°N (after Marshunova).
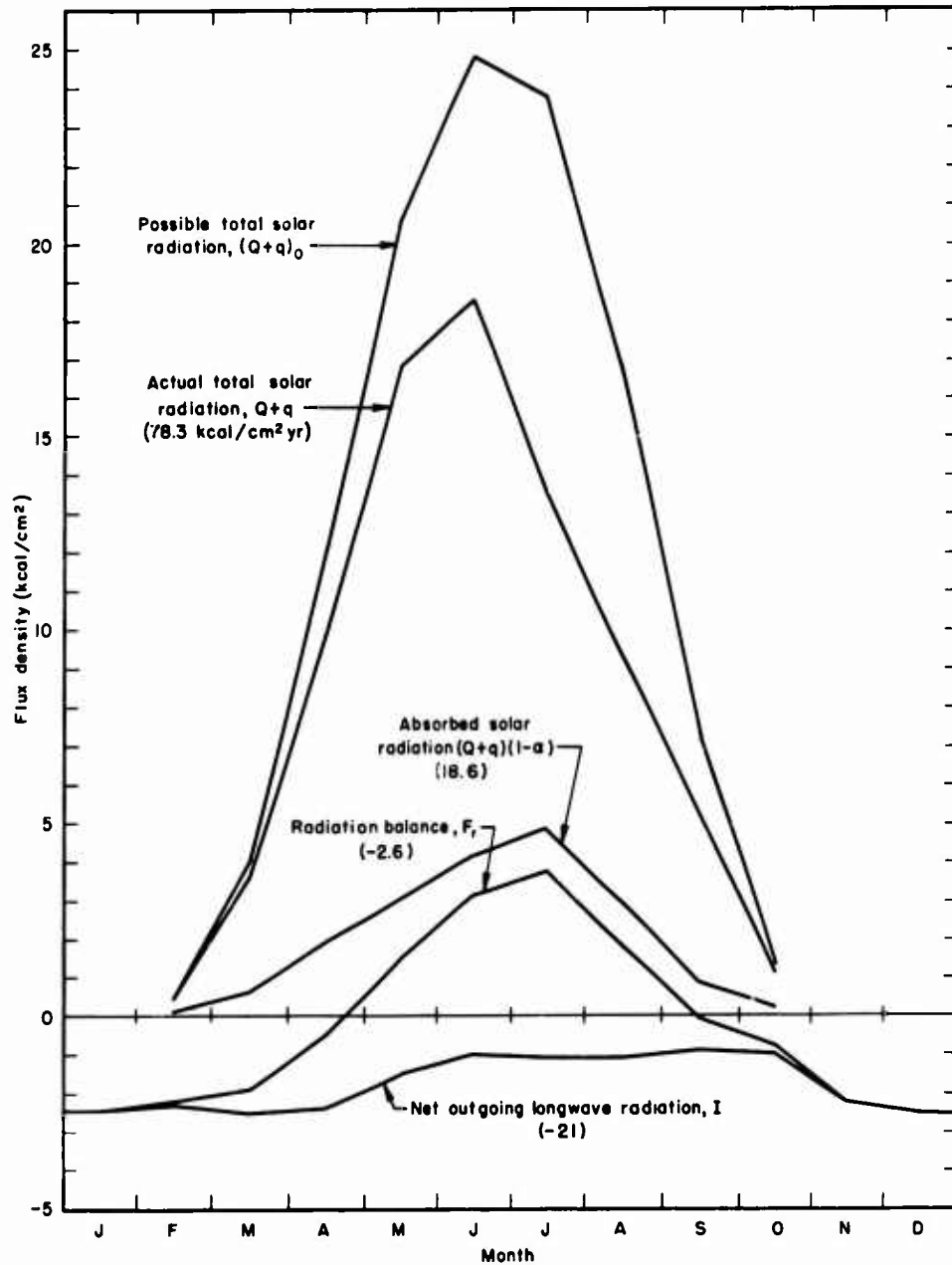
Fig. 6 -- Components of the radiation balance at the
surface at 75°N, 180°E. (The transparency
of this figure, included in the back of
this Report, facilitates comparison
with Figs. 5 and 7.)
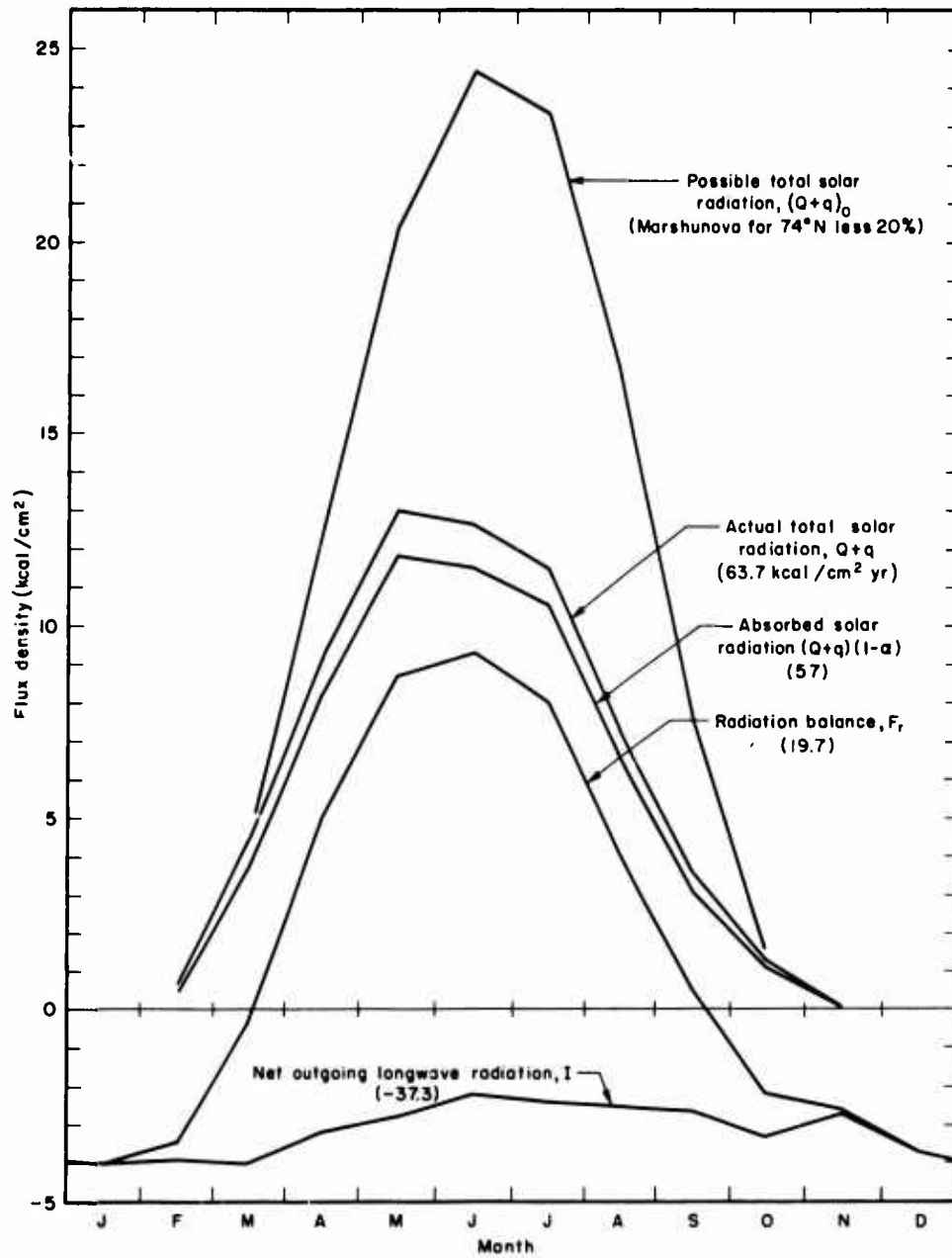
Fig. 7 -- Components of the radiation balance at the surface
in the ice-free areas of the Barents Sea.

when temperature inversions are most extreme, the maximum downward
flux occurs not at the surface but at a height of about 1 km.

The curves of net upward flux vs. calendar time parallel the
values at the surface up to 3--4 km, but above that level they tend
to be stable through the year. Of course, the presence of cloudiness
changes these patterns greatly.

The alteration due to clouds is chiefly confined to lower levels,
for most cloudiness in the Arctic is at considerably lower levels and
of less vertical extent than in other areas. Most Arctic clouds are
below 500 m. According to Soviet aerial reconnaissance (Marshunova),
the mean heights of clouds are:

Cloud Type	Bottoms	Tops
Stratus	200 m	660 m
Stratocumulus	900 m	1300 m
Altostratus	250 m	3200 m
Altocumulus	2700 m	3100 m

But the same source gives the mean thickness for stratus as 350 m and
for stratocumulus, 540 m. These cloud types predominate throughout
the summer.

Figure 4 shows how cloudiness alters the values of the longwave
fluxes at the surface. The net longwave loss at the surface is in
fact minimum in summer rather than maximum. Under a low overcast
(typical summer) the longwave balance is close to zero.

At the top of the cloud layer (660 m for summer stratus), there
is a sharp increase in longwave-radiation intensity for 1--2 km,
above which the change is slower. At higher levels the annual changes
are weaker when clouds prevail.

RATE OF COOLING OF THE ATMOSPHERE

Ordinarily the radiational cooling in a clear atmosphere is
relatively uniform throughout the atmospheric column (Fig. 8). In
the central Arctic, however, the intensity of the winter temperature
inversions alters this picture.

Figure 9 shows changes in radiational cooling with altitude for
January and July at four stations that typify regional conditions.
On the left is shown the corresponding temperature stratification.
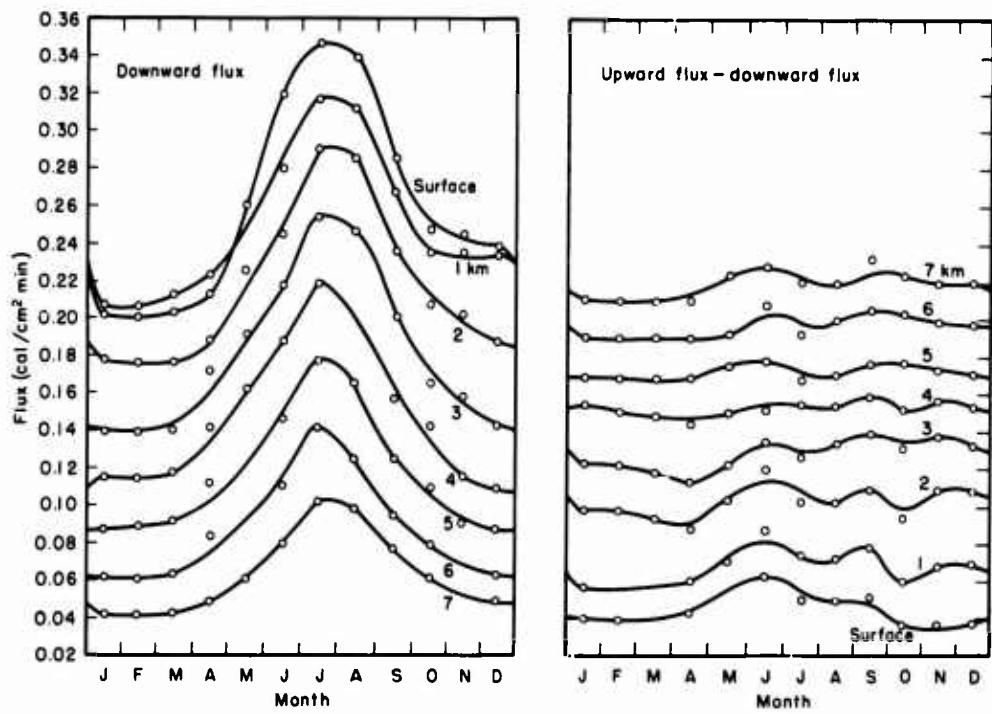
Fig. 8 -- Longwave-radiation fluxes in a clear atmosphere
in the central Arctic (after Marshunova).

Through most of the troposphere, radiational cooling amounts to 1.0--1.5°C/day, but in the central Arctic in winter the lower 1--2 km may be cooled at 2.0--2.5°C/day. This layer of rapid radiational cooling corresponds to the warmest layer of the atmosphere, which is being cooled from below by turbulent heat exchange with the surface. If clouds are present, the cooling near their upper surface can even be 3.0--3.5°C/day.

THE RADIATION BALANCE OF THE EARTH--ATMOSPHERE SYSTEM

The radiation balance of the earth--atmosphere system was represented by Eq. (3) as:

$$B_{e/a} = (Q + q)(1 - \alpha) + q' - I_{\infty} \quad .$$

The first component, $(Q + q)(1 - \alpha)$, is the solar radiation absorbed at the surface. Values of this component for grid points in the Arctic are given in Table 4.

The second component, q', is the solar radiation absorbed directly by the atmosphere, and I_{∞} is the outgoing longwave radiation to space. Appropriate values for these components must be found in order to determine the radiation balance for the earth--atmosphere system as a whole. These latter two components will be discussed separately before considering the composite picture.

Absorption of Solar Radiation by the Atmosphere, q'

The largest quantitative uncertainty in the heat budget lies with the amount of solar radiation absorbed directly by the atmosphere (including clouds). Absorption by water vapor and by CO_2 is well known and can be calculated with acceptable precision. On the other hand, absorption due to clouds and turbidity in the atmosphere is very poorly known. This uncertainty is due both to uncertain knowledge of the amounts of clouds and turbidity in the atmosphere and to poor understanding of the effects of a given condition on solar radiation.

The extent and type of summer cloudiness has already been discussed. The typical summer-cloud condition is stratus and strato-cumulus, which persists from about 70 per cent to 90 per cent in time
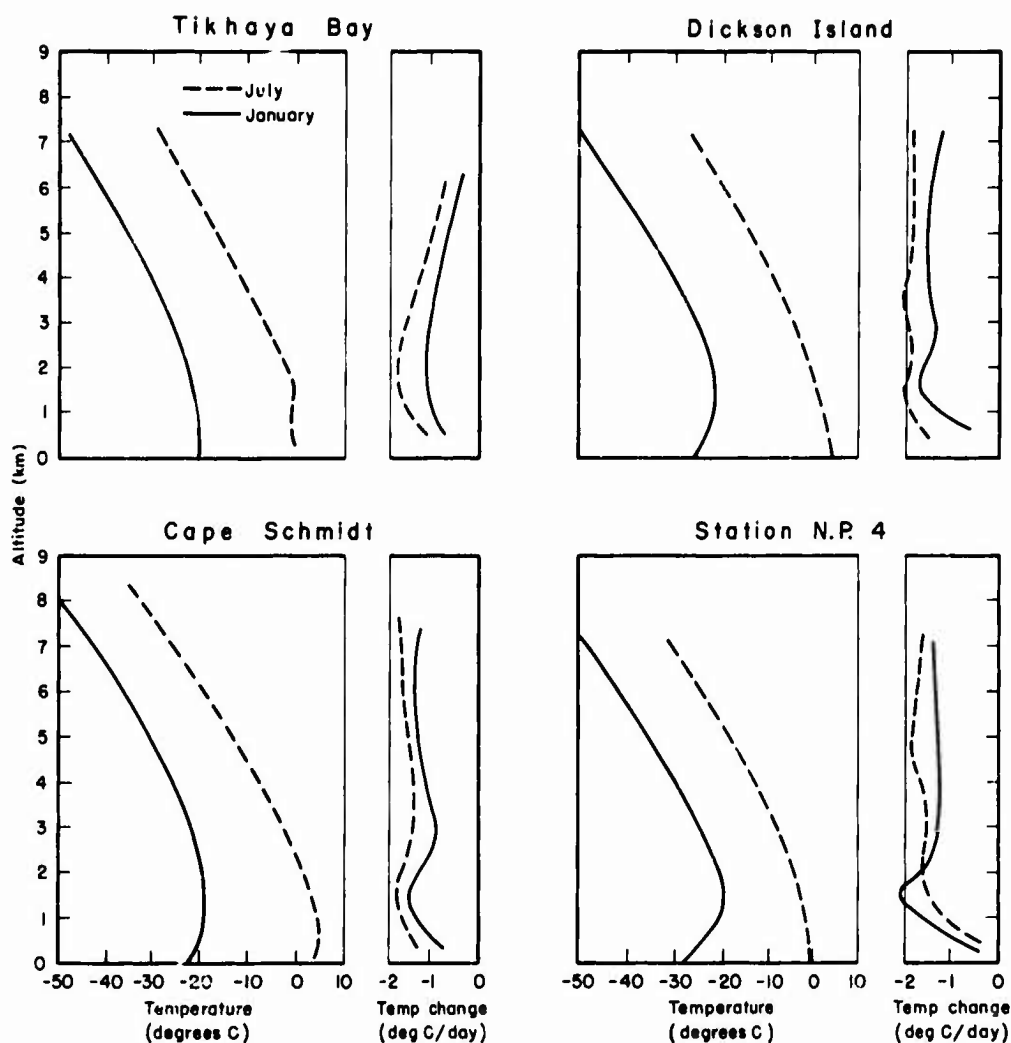
Fig. 9 -- Change in radiational cooling of the atmosphere with
height for January and July. The two left curves
for each area show temperature stratification of
the atmosphere; the right-hand pairs show
cooling rate.

and space over the Arctic Basin. Higher clouds are much less prevalent. The typical thickness of each layer is about 0.5 km.[3] This cloud condition can be expected to affect greatly the pattern and amount of solar-radiation absorption.

Probably nowhere else on earth does such a simple typical cloud structure persist over such great areas so much of the time. However, relatively little attention has been given to field measurements of either cloud characteristics or solar absorption. The only group known to have attempted such measurements is that of I. M. Dolgin, head of the Climatology Laboratory at the Arctic Institute in Leningrad. The results of this work are not yet available. However, in an earlier paper,[13] Dolgin and others report an annual value of 50 kcal/cm^2 for solar radiation absorbed by the atmosphere in the central Arctic, out of 135 kcal/cm^2 incoming at the top of the atmosphere. The corresponding figure given by Dolgin, et al., for Antarctica is only 25. Thus the high figure for the Arctic must be associated with the prevalence of summer stratus, which lets through most of the incident radiation and tends to trap solar radiation between the high-albedo ice surface and the highly scattering cloud layer.

The absorption percentage reported by Dolgin is about twice as large as has usually been assumed for mid latitudes.[14] However, the author decided to accept his value for this report, and it is reflected in Fig. 10.

The justification for this choice is based on a review of the rather meager amount of related data as follows:

The very few field measurements available are reported in Refs. 15 and 16. Neiburger's measurements were on California stratus (1948) and Fritz's were in deep clouds. In both cases the reliability of the results suffers both from lack of information about basic cloud parameters at the time of observation and from large quantitative uncertainties arising from the fact that the absorbed radiation is obtained as the difference between relatively large and nearly equal measured quantities. For clouds of about 0.5 km thickness Neiburger
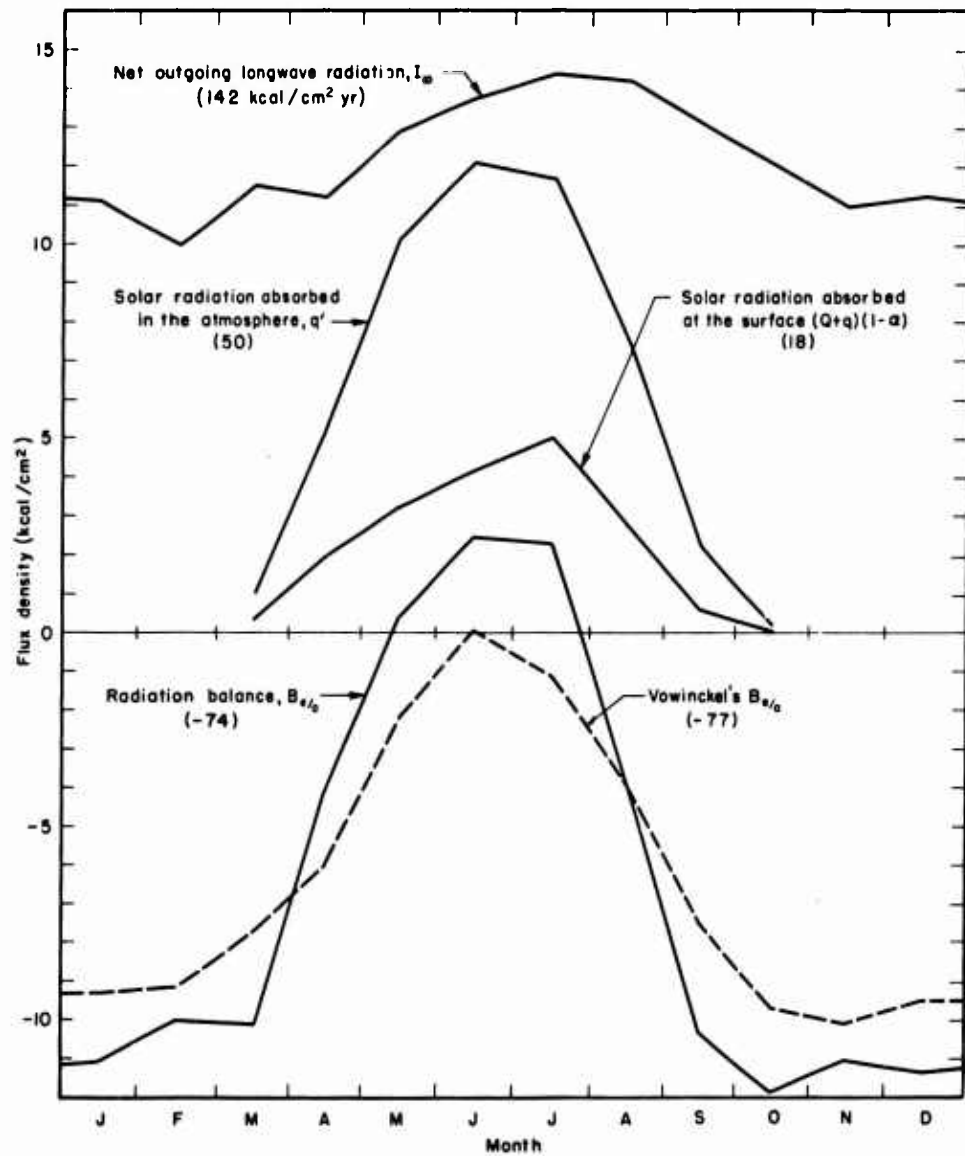
Fig. 10 -- Radiation budget for the earth--atmosphere
system in the central Arctic with
pack ice present.

obtained absorption values of about 7 per cent of the radiation incident on the cloud; Fritz obtained values of about 15 per cent for deeper clouds.

Theoretical calculations by Fritz[17] for large drops and by Korb, et al.[18] for a variety of conditions add substantially to earlier work, but so many assumptions about basic parameters of calculations are involved that their results should be viewed with caution. (Vowinckel[6] used Korb's result plus his own assumptions about cloud conditions.)

By far the most complete theoretical treatment of the problem is by Feigelson.[19] She has carried out calculations for separate spectral bands (about 20) and for various zenith angles and cloud conditions. A rough idea of the problem can be obtained from her calculations for the following cloud model:

Solar elevation, 30°,	Water-droplet density, 5 gm/m^3,
Cloud thickness, 1 km,	Water-vapor density, 5 gm/m^3,
Cloud base, 1 km,	Extinction coefficient, 25/km,

Fraction of scattered energy

diffused forward, $\beta = 0.6$ and 0.9.

The absorbed solar radiation turned out to be (in cal/cm^2 min):

	Under Cloud $0 \leq z \leq 1$ km	In Cloud $1 \leq z \leq 2$ km	Above Cloud $2 \leq z \leq 5$ km	Upper Levels $z > 5$ km
for $\beta = 0.6$	0.0003	0.0390	0.0410	0.0610
for $\beta = 0.9$	0.0010	0.0600	0.0350	...

A significant feature of these results is that absorption of solar radiation in the atmosphere is over 50 per cent greater when the cloud is present than when the sky is clear. The total of 0.157 cal/cm^2 min (for $\beta = 0.9$) corresponds to 6.8 $kcal/cm^2$ mo, a value significantly greater than is usually assumed for solar absorption in the atmosphere.

Of course, Feigelson's model is not intended to represent conditions of Arctic summer. The solar elevation of 30° is higher than the mean for June and does not allow for diurnal variation. On the

other hand, upper clouds are not considered, nor is the effect of the high-albedo surface below, both of which might increase the value of absorbed radiation considerably. Thus we may conclude that the high value reported by Dolgin is theoretically possible.

It is also of interest to note the results of direct measurements of solar-radiation absorption in other regions. The most comprehensive measurements yet reported of all the radiative fluxes as a function of height are the balloon soundings made by Kondrat'yev, et al., in 1961--1962.[20] These measurements were taken in the Soviet Union, usually over land surfaces of relatively low albedo. Their relevance to conditions in the central Arctic can therefore be questioned, but it is significant that their measured values of absorbed solar radiation are much greater than those calculated or estimated by earlier authors.

The following table shows the values obtained theoretically by London[21] and those measured by Kondrat'yev, et al.[20]

Observer	Condition	Solar Heating (deg/hr)*	Total Solar and Long-wave Radiative Heating (deg/hr)
London:	clear cloudy	0.004--0.025 0.040--0.033	
Kondrat'yev:	clear, summer clear, winter clear, winter (in lower 500 mb layer)	0.073--0.116 0.037--0.093 0.067--0.155	0.025--0.063 0.040--0.106

*$1°$/hr for the whole atmosphere corresponds to 17.3 $kcal/cm^2/mo$.

These measurements indicate that solar-radiation absorption takes place mostly in the lower atmosphere. In the lowest 50-mb layer, Kondrat'yev's measurements of solar heating vary from $0.115°$/hr to $0.826°$/hr, depending on turbidity.

In view of the wide variation of measured values and the dependence on local conditions, it seems reasonable to conclude that the high value reported by Dolgin could easily be true. The question will not be settled until detailed measurements are made through the Arctic summer stratus. In this connection it may be noted that the balloon

apparatus used by Kondrat'yev could probably be used with a tethered kytoon to obtain radiation soundings through the Arctic summer stratus.

Longwave-Radiation Loss to Space, I_∞

Marshunova gives values of I_∞ for four typical stations in the Arctic Basin (Table 5). These values were calculated by using mean monthly values of temperature and humidity with altitude in the troposphere. The net upward flux at the tropopause was assumed to equal I_∞, that is to say, the value would not change if the isothermal stratosphere were taken into account. Mean monthly values of the height and temperature of the tropopause were taken into account, as was the difference in emissivity of clouds from a blackbody. Mean monthly cloudiness was taken into account, and the temperature of cloud tops was taken as that of the air.

The resulting mean annual values for clear skies, I_∞, was 0.260 cal/cm^2 min in winter and 0.340 cal/cm^2 min in July.

It is of interest to compare these results with those of other investigators. Table 6 shows such a comparison, and it can be seen that Marshunova agrees most closely with Houghton.

Similar calculations have been carried out by Vowinckel and Orvig.[11] They assumed blackbody radiation from clouds, used independent cloud data, and calculated values for the 300-mb level.

The relative constancy during the year of the longwave radiation to space (Table 5) can be explained by noting that this radiation originates at some depth in the atmosphere corresponding to temperature and humidity stratification (as in Fig. 8). During winter, when low temperature and humidity prevail, this level will be lower than in summer. Thus the effect of seasonal temperature change is damped by a vertical displacement (with corresponding temperature difference) of the levels from which the radiation comes.

Cloudiness also plays a role. In late winter, a significant fraction (10--20 per cent) of longwave radiation from the surface is able to pass through the atmosphere to space because clear skies prevail and atmospheric water vapor is at a minimum. During this season, however, surface temperatures (and radiation intensities) are at a minimum. During summer, when the surface is much warmer,

Table 5

ANNUAL COURSE OF RADIATION-BALANCE COMPONENTS FOR THE EARTH--ATMOSPHERE SYSTEM*

(kcal/cm² mo)

Location	Radiation Components	Month												Net Annual
		Jan	Feb	Mar	Apr	May	June	July	Aug	Sept	Oct	Nov	Dec	
90°N	Solar absorbed at surface	...	...	...	1.6	3.2	4.2	4.9	2.6	0.4	...	...	...	16.9
	Solar absorbed in atmosphere	...	...	0.1	4.7	11.0	13.0	12.2	7.5	1.5	...	...	...	50.0
	Net outgoing longwave	10.7	9.6	10.7	10.7	12.7	13.4	14.2	13.8	13.0	11.8	10.4	10.7	141.6
	Balance	-10.7	-9.6	-10.6	-4.4	+0.5	+3.8	+2.9	-3.7	-11.1	-11.8	-10.4	-10.7	-74.7
Tikhaya	Solar absorbed at surface	...	...	0.2	1.3	2.2	4.3	6.4	4.8	1.5	0.1	...	...	20.8
	Net outgoing longwave	11.2	10.0	11.2	11.2	12.8	13.6	14.5	14.3	13.0	12.4	11.2	11.4	146.8
Dickson Island	Solar absorbed at surface	...	0.1	0.6	1.5	2.9	8.1	9.5	5.1	2.1	0.2	...	...	30.1
	Net outgoing longwave	11.6	10.4	11.6	11.6	13.0	14.0	14.8	14.2	13.4	13.0	11.6	11.6	150.8
Cape Schmidt	Solar absorbed at surface	...	0.2	1.1	2.1	4.4	8.4	9.2	6.0	2.4	0.9	...	...	34.7
	Net outgoing longwave	11.5	10.3	11.4	11.8	13.2	14.2	14.6	14.6	13.4	12.4	11.6	11.6	150.6

*After Marshunova

extensive cloudiness and high concentrations of atmospheric water
vapor absorb almost all of the radiation from the surface (over 95
per cent). In this season the outgoing radiation to space originates
from higher elevations.

Table 6

MEAN ANNUAL OUTGOING LONGWAVE RADIATION ACCORDING TO VARIOUS SOURCES
$(cal/cm^2 min)$

Degrees North Latitude	Marshunova (1961)	Kondratiev and Filipovich (1952)	Simpson (1929)	Bauer and Philipps (1935)	Albrecht (1943)	Bagrow (1954)	Houghton (1954)
60--70	0.280	0.278	0.266	...	0.243	0.270	0.280
70--80	0.280	0.268	0.256	0.240	0.230	0.255	0.270·
80--90	0.270	0.260	0.252	...	0.225	0.248	0.265

The relative constancy of outgoing radiation with regard to geo-
graphical location is reflected also in Table 5. The variation in net
annual outgoing radiation with geographical location is only 6 per cent.

ANNUAL VARIATIONS IN THE HEAT BUDGET OF THE EARTH--ATMOSPHERE SYSTEM

Figure 10, which represents the radiation budget of the earth--
atmosphere system in the central Arctic, was constructed in the
following way.

The net outgoing longwave radiation to space, I_∞, is the mean of
Marshunova's values for $90°N$ and for Cape Schmidt (about $70°N$).

The solar radiation absorbed at the surface, $(Q + q)(1 - \alpha)$, is
the mean of Marshunova's values for $80°N$.

The solar radiation absorbed by the atmosphere, q', is taken to
be 50 $kcal/cm^2$ yr as reported by Dolgin, et al.[13] This is about
twice as much as reported by Vowinckel and Orvig and is without doubt
the most uncertain component in the heat budget.

For comparison, Vowinckel's corresponding values for $75°N$, $180°E$
are shown. The agreement for the earth--atmosphere system is better
than for either the earth or the atmosphere. Vowinckel's results
give greater radiation absorbed at the surface and less absorbed in

the atmosphere, with the two effects almost canceling for the earth--atmosphere system.

Figure 10 shows the annual variations of the heat budget for the earth--atmosphere system for the central Arctic. The net deficit is of special interest because it represents the amount of heat that must be supplied either to the surface from below or by the atmosphere. This quantity is important to our later discussion of the role of atmospheric advection.

In conclusion, it is instructive to consider the relative magnitudes of the components of the annual heat budget of the earth--atmosphere system in the central Arctic, according to Dolgin, et al. (Fig. 11). The total annual heat loss by outgoing longwave radiation to space is 140 $kcal/cm^2$, of which 21 $kcal/cm^2$ comes from the surface, and 119 $kcal/cm^2$ from the atmosphere.

The annual total of incoming solar radiation is 135 $kcal/cm^2$, of which 68 $kcal/cm^2$ is reflected back to space, 50 $kcal/cm^2$ is absorbed by the atmosphere,[*] and 17 $kcal/cm^2$ is absorbed at the surface. The difference between total outgoing (140 + 68) and total incoming radiation (135) is the amount of heat that must be supplied by advection (73), assuming no net heating or cooling of the region for the year. According to Dolgin, the heat budget of the surface is balanced by a transfer of 4 $kcal/cm^2$ of sensible and latent heat[**] from the atmosphere to the surface.[22] The remaining 69 $kcal/cm^2$ in the atmosphere, together with the 50 $kcal/cm^2$ absorbed solar radiation, make up the 119 $kcal/cm^2$ of longwave-radiation loss by the atmosphere. For comparison, Fig. 12 shows similar components for the earth as a whole according to Budyko[22] and Houghton.[14] It should be noted that, for the earth as a whole, the percentage of solar radiation absorbed by the atmosphere is less than half of the corresponding percentage given for the central Arctic.

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[*] Presumably about 5 per cent in the ozone layer.

[**] We will later revise this picture to reflect 3 $kcal/cm^2$ yr latent heat flow from the surface to the atmosphere. Otherwise Dolgin's annual values are in good agreement with the conclusions of this study.
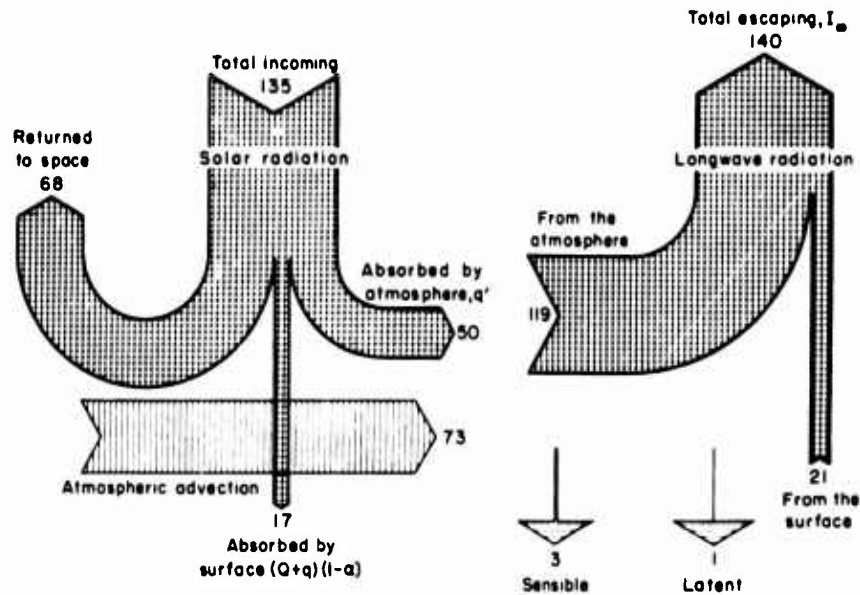
Fig. 11 -- Annual heat budget of the earth--atmosphere system in the central Arctic, in kcal/cm^2 yr (after Dolgin). Oceanic heat and melting of snow are assumed to balance and are omitted.
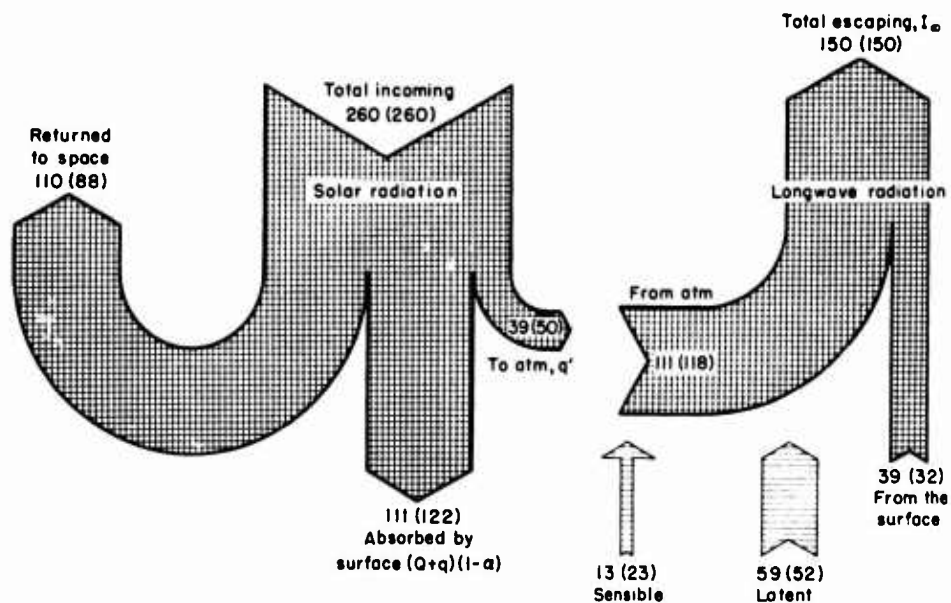
Fig. 12 -- Annual heat budget of the earth--atmosphere system for the entire planet, in kcal/cm^2 yr (after Budyko, 1962). Parentheses show corresponding Northern Hemisphere values by Houghton (1954).

III. TIME AND SPACE VARIATIONS IN THE RADIATION CLIMATE

Three questions of special interest in the subsequent discussion are:

1. What is the most realistic representation for the mean radiation climate of the Arctic Basin?

2. How much do year-to-year values vary from mean conditions and what physical factors account for such variations?

3. What would the radiation climate be if the pack ice were removed?

THE MEAN RADIATION CLIMATE FOR THE ICE-COVERED AREA
OF THE ARCTIC BASIN

As noted in Section II, the two regional factors that basically determine the Arctic's radiation climate are the character of the surface, and the cloud conditions. If these factors are the same in two regions, then differences in latitude or longitude are of relatively minor importance. Comparison of Figs. 5, 6, and 7 show this dependence clearly.

The surface consists of alternating pack ice and open leads, but since open leads represent such a small fraction of the total area (less than 1 per cent), the mean radiative character of the surface cannot vary much from place to place for that reason. In all seasons, the radiative characteristics of the pack ice are relatively uniform over the whole area it covers. The largest variation occurs in mid summer, when puddles appear earlier at lower latitudes. The maximum effect of this variation should be approximated by comparing Figs. 5 and 6, which are separated by 15° of latitude. As can be seen, the differences in net radiation are small. According to an independent study by Chernigovskiy the absorbed solar radiation over pack ice at

75--78°N is about 10 per cent greater than at 90°N.[23] This checks almost exactly with Fig. 5 and Fig. 6.

The variation in cloudiness from place to place is significant, with somewhat greater cloudiness being prevalent in the European sector than in the Greenland and American sectors. According to Gavrilova this difference accounts for somewhat larger quantities of total radiation, Q + q, in the American sector (70--85 kcal/cm^2 yr) than in the European sector (60--70 kcal/cm^2 yr).

In view of the foregoing, Fig. 13 was constructed by choosing 80°N as a representative mean latitude for the pack-ice area and averaging the values for the four different longitudes at 80°N given in the Appendix. In this way variations of albedo with latitude and cloudiness with sector are compensated for. Figure 13, therefore, represents the approximate mean radiation climate for the pack-ice area according to Marshunova's data. It may be noted that the value of 18 kcal/cm^2 for absorbed radiation agrees well with the value of 17 given by Dolgin (Fig. 11) and that 20.7 kcal/cm^2 net outgoing longwave radiation agrees well with Dolgin's value of 21.

A slightly higher annual value of radiation balance has been obtained by Chernigovskiy, using the mean results of seven Soviet drifting stations (Fig. 14).[24] By comparing Fig. 13 with Fig. 14 it is apparent that the difference is due to Marshunova's somewhat larger net outgoing longwave radiation during the winter months. Since average winter cloudiness is probably somewhat greater in the Eurasian sector than in the North American sector, it is possible that Fig. 13 is truly representative of the whole central Arctic and Fig. 14 representative of the areas of Soviet drifting stations. It is also possible that, in computing back radiation from the atmosphere during winter, Marshunova is underestimating the turbidity or cloudiness of the atmosphere. On the basis of present knowledge the author concludes that either Fig. 13 or Fig. 14 could be accepted as the mean radiation climate of the central Arctic. The true mean values probably lie somewhere between these two representations.[*]

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[*]Vowinckel and Orvig, however, report much more positive values, ranging from +7 kcal/cm^2 up, at 90°N. For reasons given later the author concludes that these values are too high and probably are due to underestimates of surface albedo.
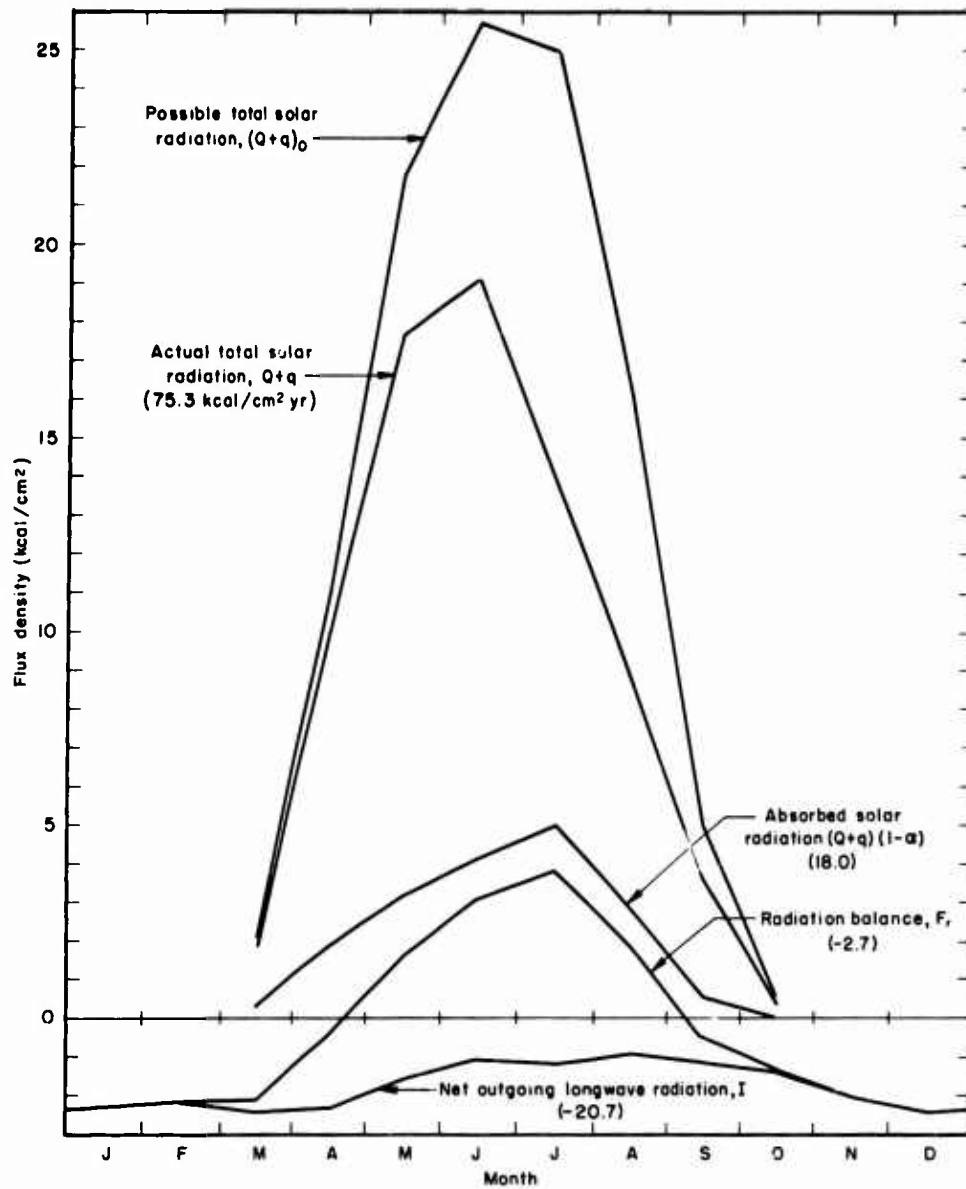
Fig. 13 -- Estimated present radiation budget at the
surface in the central Arctic.

INTERACTION OF SURFACE ALBEDO, CLOUD COVER AND RADIATION BALANCE

The radiation balance of the surface is sensitive to changes in albedo or cloud cover, and during the critical summer months, albedo depends on radiation balance, because the onset and extent of puddling alters the amount of solar radiation absorbed. Thus anomalies in this interaction can cause anomalies in the heat budget as a whole.

Figure 15 shows the interdependence of these factors according to Marshunova (assuming cloud types and temperatures characteristic of the central Arctic).

In Fig. 15 it can be seen that at all solar altitudes the radiation balance for snow is higher with overcast than with clear skies. That is, if solar radiation is scattered, more of it is absorbed than when only direct radiation strikes the snow surface.

A similar relationship holds for a water surface at low solar angles, but above 8°--10° solar height the opposite is true. At mid summer in the central Arctic (say solar height 30°) the radiation balance for water is about twice as great under clear skies as under overcast conditions.

Under low overcast, the radiation balance for the snow-covered pack ice is positive during the whole light period, passing through zero at solar height 0--2°. This is due to the combined effect of less longwave loss, I, and more solar absorption at the surface, $(Q + q)(1 - \alpha)$.

Under clear skies the radiation balance becomes positive for a water surface at 6° and for snow cover at 15° solar height.

These relationships are vital to assessing the effects of artificially altering surface albedo or cloud conditions. They also give some insight into possible physical causes of seasonal anomalies. For example, it is significant that, over snow, the presence of Arctic stratus not only increases the solar heat absorbed by the atmosphere, but also increases the positive radiation balance at the surface. Thus cloud dissipation during May would make the heat budget of the surface more negative. During July, however, when mean surface albedo is lower, cloud dissipation could make the heat budget of the surface more positive.
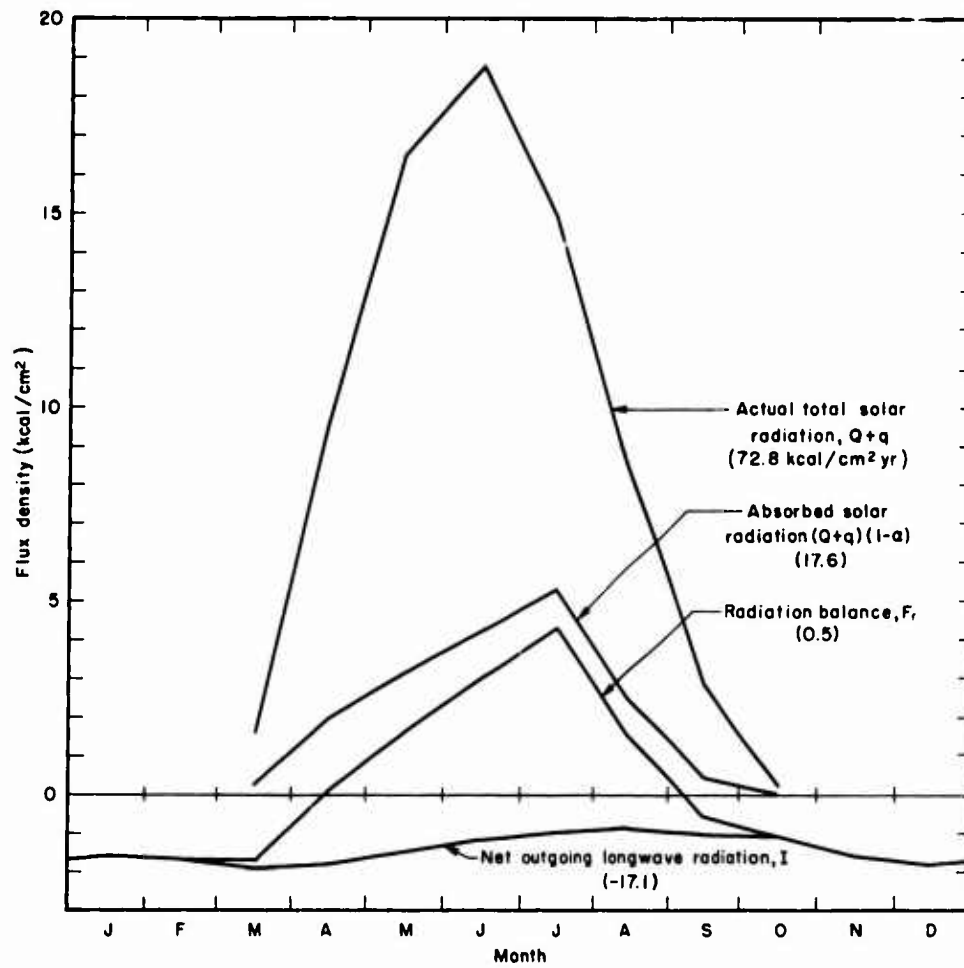
Fig. 14 -- Radiation components at the surface in the
central Arctic (after Chernigovskiy).

YEAR-TO-YEAR VARIATIONS IN CONDITIONS IN THE CENTRAL ARCTIC

In view of the interdependence just discussed it is not surprising that year-to-year variations in the radiation budget are found to be substantial. During some years (1959, 1961) Atlantic cyclone tracks tend to follow more northerly courses during May and June than in other years (1960). The variation in cloudiness and snowfall in the central Arctic during this critical period could appreciably influence the radiation budget.

Should abnormal conditions in early summer advance or delay the date of puddling on the ice, the consequent change in surface albedo is significant. During the summer months total solar radiation at the surface reaches 18--20 kcal/cm^2 mo, so a 20 per cent change in albedo for even one month can mean 3--4 kcal/cm^2 to the surface radiation balance.

On the basis of Soviet field measurements of radiation components and related meteorological parameters, Gavrilova estimates year-to-year variation in total solar radiation at the surface, $Q + q$, to be 8--10 per cent, variations in net outgoing longwave radiation at the surface to be 20 per cent and variations in the radiation balance at the surface to be from 10 per cent year-to-year in summer to 50 per cent during transition seasons.

During winter and spring, albedo varies little from place to place and from year to year, but in summer, anomalies may occur. Chernigovskiy reports that on days with summer snowfall the albedo fluctuated from 98 per cent (for fresh snow) to 21 per cent (for melted snow water), a total change of 77 per cent. [24]

Since 1954 there have always been two or more Soviet drifting stations taking complete actinometric observations. During this period the annual radiation balance, F_r, has been observed to vary substantially. The highest monthly value observed was 7.4 kcal/cm^2 for July 1955. The average annual value is around zero, but in 1955--56 North Pole 4 recorded 5.5 kcal/cm^2, and in 1957--58 North Pole 6 recorded 3.8 kcal/cm^2. [24]

From these data it would appear that year-to-year variations are significantly larger than the uncertainty of mean values. This
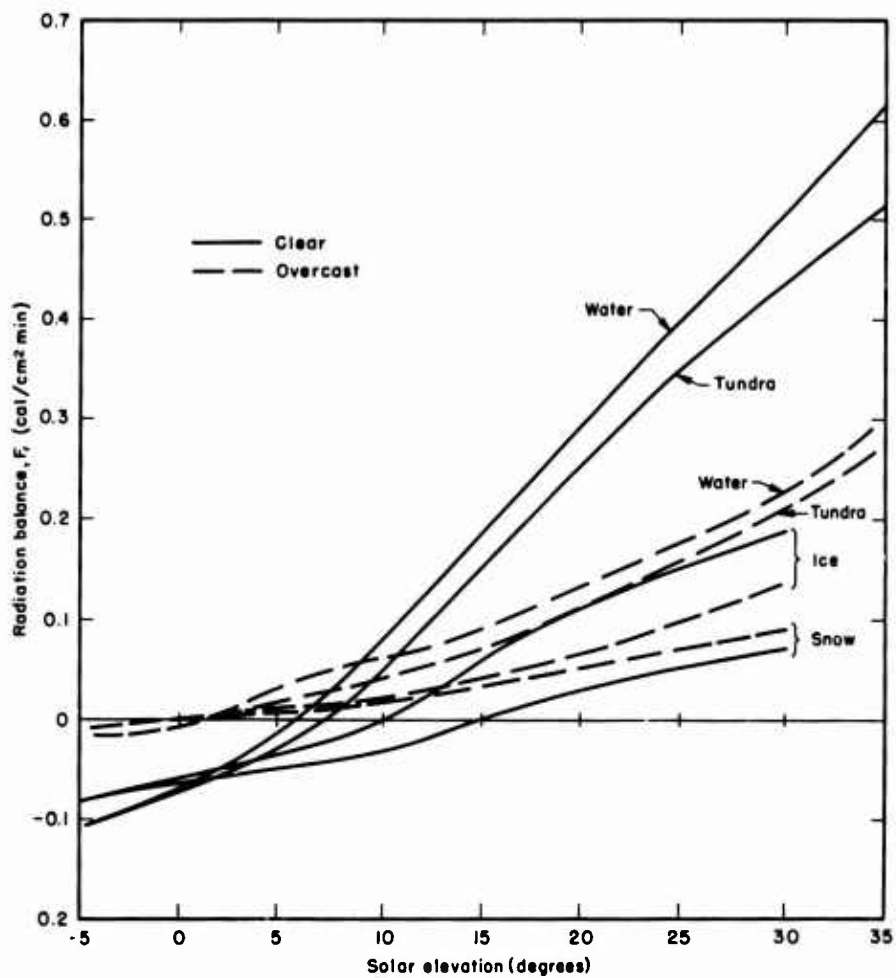
Fig. 15 -- Radiation balance vs. solar height for two
conditions of cloud and four of surface.

important conclusion will be referred to later in connection with future studies of heat budget and climatic anomalies.

THE RADIATION CLIMATE OF PERIPHERAL SEAS

Figure 7 is a good representation of the radiation climate of peripheral sea areas that are largely free of pack ice. Cloudiness in the Barents Sea is especially great because it is in the path of many Atlantic cyclones. In ice-free areas of less cloudiness, such as the Chukchi Sea, the annual radiation balance may be some-- what higher. Recent reports[25] from scientists at the Main Geo-physical Observatory, Leningrad give monthly values as high as 10 $kcal/cm^2$ for such areas (slightly higher than in Fig. 7). Cherni-govskiy reports that absorbed radiation over pack ice in the peri-pheral zone $70^{\circ}N$--$75^{\circ}N$ is 22 per cent greater than at $90^{\circ}N$.

THE RADIATION CLIMATE OF AN ICE-FREE CENTRAL ARCTIC

As noted earlier, Fig. 7 can be regarded as a first approxi-mation to an estimated radiation climate for an ice-free Arctic. However, the Barents Sea is at too low a latitude to be truly rep-resentative of the central Arctic, and its location on the Atlantic cyclone track reduces the available solar radiation at the surface to an abnormally low value. The annual radiation balance of about 20 $kcal/cm^2$ yr may be lower than would be typical for an ice-free central Arctic.

Another estimate can be produced by adjusting the data of Fig. 13 according to the observed behavior of the radiation pro-cesses. Figure 16 is the result of such an adjustment and was constructed as follows:

Possible total solar radiation at the surface was estimated by reducing the value of Fig. 13 by 20 per cent to allow for the loss of diffuse radiation caused by the reduction of surface al-bedo.

Actual total solar radiation at the surface was obtained by assuming 25 per cent diminution of the possible total by clouds.
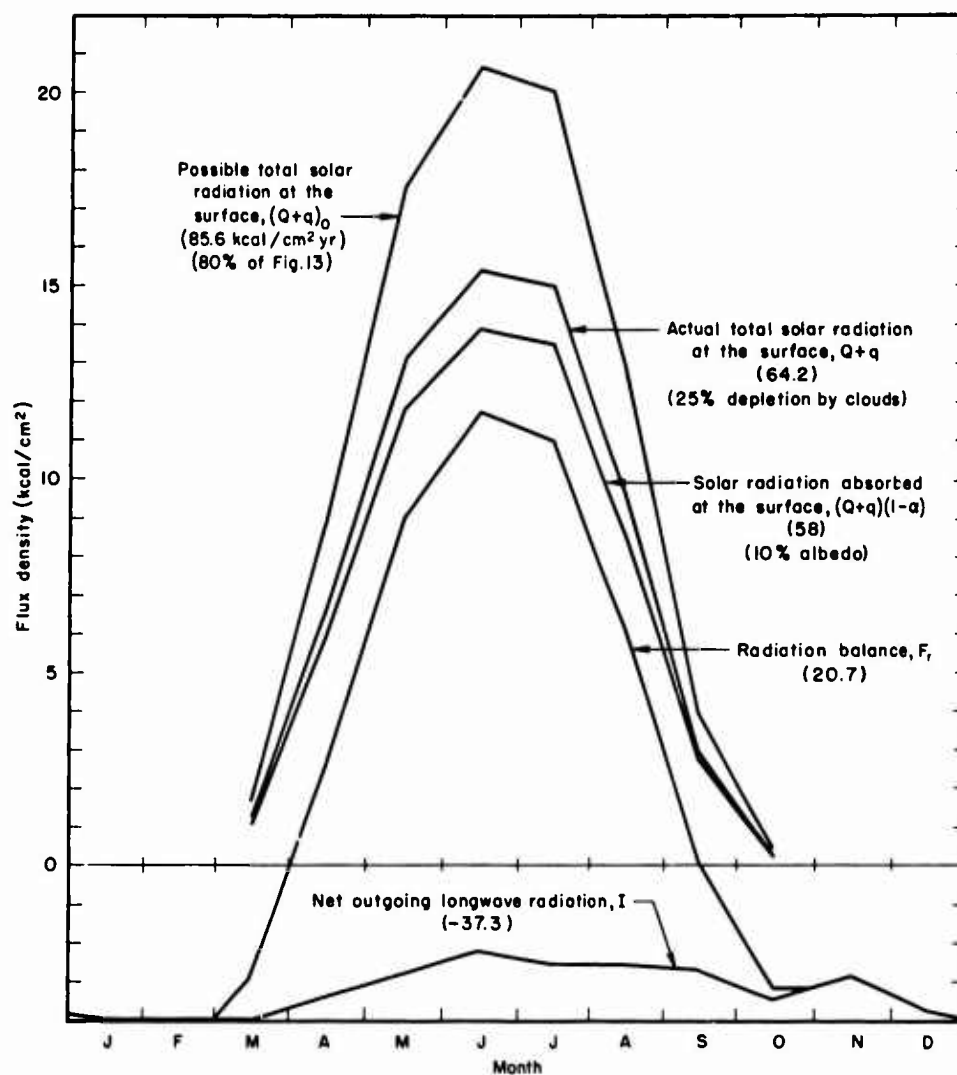
Fig. 16 -- Estimated radiation budget at the surface
for an ice-free central Arctic.

Albedo of the water surface was assumed to be 10 per cent, a reasonable value for diffuse radiation.

Net outgoing longwave radiation was taken as the same as in Fig. 7. This assumption carries the inference that water temperature (but not necessarily the vertical cloud extent) and mean horizontal extent of cloudiness would be similar to present conditions in the Barents Sea.

The result is a radiation balance of 21 $kcal/cm^2$ yr, compared to 20 $kcal/cm^2$ yr for Fig. 7.

Although we can only guess about atmospheric circulation and mean cloud conditions for an ice-free Arctic, Fig. 16 is probably a reasonable estimate of a mean radiation budget, and it will be used in the remainder of this study.

RADIATION BUDGET OF EARTH--ATMOSPHERE SYSTEM FOR AN ICE-FREE ARCTIC

Figure 17, which represents the estimated radiation budget of the earth--atmosphere system for an ice-free Arctic, was constructed in the following way.

It is assumed that the net longwave radiation to space, I_∞, would be about the same in summer as for the "ice-in" condition (Fig. 10); in winter it would be higher. To estimate how much higher, Vowinckel's grid-point data for $75^{\circ}N$, $180^{\circ}E$ (an ice-in point near the Chukchi Sea) were compared with his $75^{\circ}N$, $0^{\circ}E$ (an ice-free point in the Greenland Sea). The difference is about 1.2 $kcal/cm^2$ mo in winter. Thus the outgoing radiation from Fig. 10 was raised by a corresponding amount. Of course, such a procedure implies cloud conditions and temperature--humidity structure similar to th·t now typical of the Norwegian Sea. What actual cloud conditions would really be like over an ice-free Arctic Ocean is at present a speculation, but in Section VI, arguments will be given that it would be quite cloudy in winter. For the present, the above estimate seems a reasonable approximation for the purpose of establishing an over-all view of the heat budget for ice-free conditions.

The solar radiation absorbed at the surface, $(Q + q)(1 - \alpha)$, was taken from the estimated surface radiation budget shown in Fig. 16.
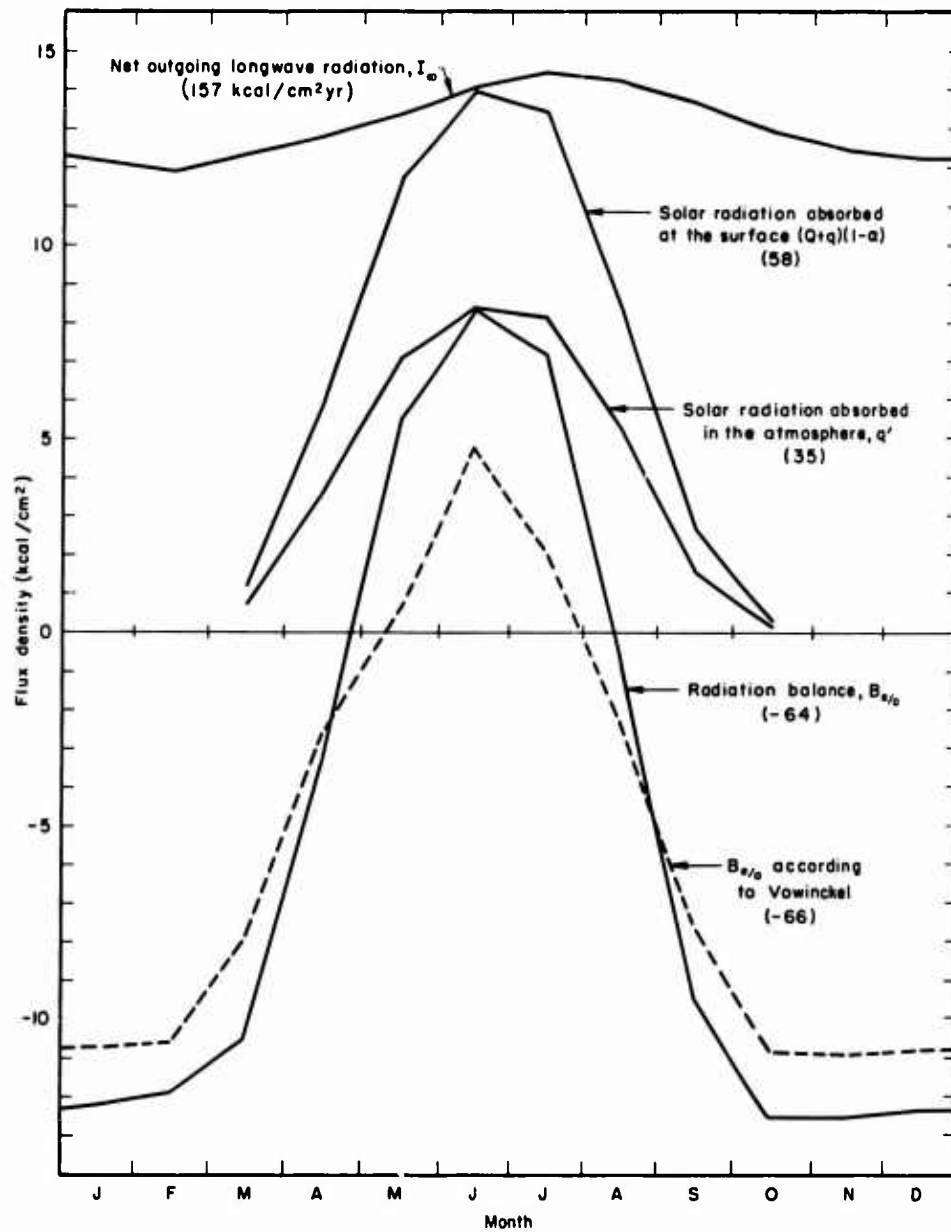
Fig. 17 -- Estimated radiation budget for an earth--atmosphere
system with an ice-free central Arctic.
(Vowinckel's data are for 75°N, 0°E.)

The solar radiation absorbed by the atmosphere was taken arbitrarily as 35 kcal/cm^2 yr, which is a somewhat greater percentage than that given by Vowinckel and Orvig for the Arctic or that by Houghton for the Northern Hemisphere. (Vowinckel used Houghton's depletion coefficients.) On the other hand, it is lower than reported by Dolgin, et al., for the present central Arctic. The percentage should be lower because the low albedo of the water surface reduces the solar-radiation intensity in the lower atmosphere by about 20 per cent under the stratus layer typical of summer ice-in conditions. It is doubtful whether such a stratus layer, which acts as a sort of solar-radiation trap, would be as widespread or persistent over an ice-free ocean; on the basis of this rationale, a reduction from 50 to 35 kcal/cm^2 in annual solar radiation absorbed by the atmosphere has been made from ice-in to ice-free conditions.

The radiation balance of the system is the resultant of the above three curves. For comparison, Vowinckel's corresponding values for 75°N, 0°E are shown.

FUTURE FIELD OBSERVATIONS OF THE ARCTIC RADIATION CLIMATE

From the foregoing discussion, some general observations can be made about the kinds of additional field observations that are sorely needed.

The first and most general comment is that the very success of past research, which tried to describe the basic characteristics of the Arctic radiation climate, now makes it possible to shift the emphasis toward studies of the interaction of the Arctic heat budget with atmospheric circulation and the climates of other regions. The year-to-year variations in the surface radiation balance are probably greater than our uncertainty about mean values. Observing these variations should yield new insights into the complicated interaction with the atmosphere while simultaneously reducing uncertainties about mean conditions.

Solar-Radiation Measurements

Improvement in understanding the solar-radiation components hinges on better understanding of the transformation of incoming solar radiation by clouds and at the earth's surface. The reflectivity and scattering properties of Arctic clouds are poorly known, as are the extent, the height, and the type of cloudiness. Changes in mean surface albedo have tremendous impact on the heat budget, but have received relatively little attention in field-observation programs.

Arctic-Cloud Conditions

It is ironic that after nearly twenty years of routine United States weather-reconnaissance flights over the Arctic, the data obtained have never been reduced to mean cloud conditions. In fact, even the elaborate computations carried out by McGill University scientists were not based on this data source. Their cloud maps for the Arctic are largely based on surface observations at land stations, a very unsatisfactory basis for drawing cloud maps for the pack-ice area. One simple and very useful contribution to knowledge would be to translate the cloud data from past aerial reconnaissance to a more usable form. At the same time the data from current and future reconnaissance should be reviewed for usability and arrangements made for its optimum utilization.

Physical Properties of Arctic Clouds

The reflecting and scattering properties of Arctic clouds have never been systematically investigated, but important assumptions about these factors are made in all calculations of solar-radiation components. An instrumented aircraft and a well planned summer program could investigate the validity of these assumptions.* As noted in connection with Fig. 10, the annual solar radiation absorbed in the atmosphere as given by Dolgin, et al., is about three times the value computed by Vowinckel. This is a good index to present uncertainties.

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*The instrumented C-130 operated by the Air Force Cambridge Research Laboratories and the University of California Visibility Research Laboratory might be appropriate for such a program.

Surface Albedo as a Function of Season and Location

No other single factor is likely to be more important to observe on a year-to-year basis. Yet during and since the International Geophysical Year there has been no systematic effort by the United States to obtain such observations. United States weather-reconnaissance programs should be reviewed with the aim of observing changes in surface albedo over the pack ice during the critical months of June, July, and August each year. The increasing prospect of relating heat-budget anomalies to atmospheric-circulation patterns should lend urgency to this particular program.

Longwave Radiation Measurements

Mean Cloudiness. Since cloud conditions influence the longwave-radiation balance at the surface more than does any other factor, the foregoing emphasis on observing mean cloud conditions applies during winter as well as summer.

Properties of Clouds. Marshunova reports that the emissivity of Arctic clouds at $-24°$ is 0.85, and most Soviet data are based on such an assumption. Vowinckel, et al., at McGill University, reject this assertion and assume that clouds (other than cirrus) radiate as black bodies. The effect of this assumption on the computed longwave balance at the surface is substantial. The fact is that little is known about the radiating properties of clouds at low temperatures. This uncertainty could be removed by systematic measurements.

Surface Measurements. Although the longwave and solar components are comparable in magnitude, the measurement program conducted by the United States has emphasized solar radiation. The instruments at United States stations should be augmented to permit complete longwave as well as solar measurements at the surface.[*]

Vertical Distribution. The vertical variation of the flux of long-wave radiation and its relation to temperature and humidity stratification is basic to understanding radiational cooling in the atmosphere.

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[*]For a description of instruments see Ref. 26. At United States stations Beckman and Whitley, Agmet and Schultz radiometers have been used -- but not consistently. A comparison of both solar and longwave instruments with Soviet counterparts would also improve the value of the data.

So far, the relationships have been mostly computed, and the computations involve uncertain assumptions. Direct measurements of these relationships can be made by balloon[20] or by instrumented aircraft. Such an investigation should help in the interpretation of infrared-radiation measurements from satellites.

Satellite Measurements

Satellites in polar orbits, such as Tiros and Nimbus, should, in time, make it possible to monitor the Arctic's radiation budget for indefinitely extended periods. In the meantime, however, much must be learned about how to interpret such measurements.

Nimbus I (September 1964) was the first observatory of this type in a polar orbit. It operated about a month and included visual and high-resolution infrared (HRIR) observations. The results indicate that the extent and approximate temperature of the highest cloud layer can be observed both day and night by the HRIR system. No measurements were made of outgoing radiation, I_∞. If this type of observation is combined with simultaneous aircraft measurements of radiation components in the troposphere, it should become possible to better interpret satellite measurements in terms of radiation-budget components.

IV. THE HEAT BUDGET AT THE SURFACE OVER PACK ICE

Radiation plays the major role in the heat processes taking place at the surface. Not only does radiation account for the largest turnover of energy, but also its annual variation largely determines the pattern and magnitude of the other processes. During the winter the surface characteristically loses heat by radiation and is cooled until this loss is balanced by heat from the atmosphere (turbulent exchange) and heat from below (conduction). During the summer the surface gains heat by radiation and loses heat to the atmosphere (by evaporation and turbulent exchange) and to the ice (by melting and warming).

The heat balance at the surface can be described by equating all the fluxes of heat to and from the surface as follows[*] (all fluxes _to_ the surface taken as positive):

$$F_r + F_s + F_\ell + F_c + F_m = 0 \; , \tag{5}$$

where F_r = radiation balance at the surface,

F_s = flux of sensible heat to the surface by turbulent exchange with the atmosphere,

F_ℓ = flux of latent heat by evaporation or condensation,

F_c = flux of heat by conduction to the surface from below, and

F_m = flux of heat utilized in melting (negative) or freezing (positive) of ice.

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[*]This notation follows Untersteiner and seems least ambiguous. Soviet notation is normally $R = LE + p + A$ where the terms represent radiation balance, evaporation, turbulent exchange, and exchange with deeper layers.

To understand the heat budget at the surface, we shall now assess each of these components and their interaction during the year. It will be seen that the most serious uncertainties lie with the sensible and latent heat exchange between the surface and the atmosphere, F_s and F_ℓ , where the assessments of different authorities differ widely.

AUTHORITATIVE ESTIMATES OF THE HEAT BUDGET OF THE SURFACE

Investigation of the Arctic heat budget has been the main theme of Soviet Arctic research during the past decade. The concerted effort to determine the quantitative values of the heat components began with the Soviet expedition North Pole 2 in 1950--51.[*] The lessons learned from North Pole 2 were then used in a larger and more thorough effort beginning in April 1954 with North Pole 3 and North Pole 4. Up to the present time there have been more than a dozen Soviet drifting stations and hundreds of temporary stations supported by aircraft.

The first attempts to put together all the components of the heat budget were made by Yakovlev,[27] based initially on data from North Pole 2, but later augmented by data from North Pole 3 and North Pole 4. His findings are summarized in Table 7. He obtained the heat flux by evaporation as the residual component in the heat budget; the uncertainties in the other components were thus accumulated in his values. Subsequent research indicates that the radiation balance is near zero or negative for the central Arctic. If this value is substituted for Yakovlev's radiation balance, his annual heat loss by evaporation would be reduced by about 3 $kcal/cm^2$.

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[*]Announcement of the existence of this station, which most of the time drifted in the sector north of Alaska did not come until 1954. Data were released in 1956 in connection with IGY planning.

Table 7

COMPONENTS OF THE HEAT BUDGET OF THE SURFACE ACCORDING TO YAKOVLEV
$(\text{kcal}/\text{cm}^2)$

Component	Sum: May—Aug	Annual Sum
Radiation balance, F_r	11.0	2.9
Heat of evaporation, F_e	-5.7	-7.0
Heat of melting, F_m	-2.9	-2.8
Turbulent heat exchange, F_s	-2.2	1.5
Warming of the ice $\left.\right\}F_c\left\{\right.$	-1.5	0
Heat from the ocean	0.9	5.5

There have been four United States drifting stations;
heat-budget investigations have been prominent in their programs,
although observations have not been so elaborate as in the Soviet's.
Most of this work, performed by scientists of the University of
Washington, has been reported by Badgley,[28][29] whose results
are summarized in Table 8.

The group at McGill University, led by E. Vowinckel[30][31]
and Svenn Orvig, had published 13 heat-balance studies as of December
1964. They arrived at estimates by drawing maps of the basic
meteorological parameters and then calculating each of the heat
fluxes for grid points in the Arctic.

Unfortunately, some of the disagreements among these sources
are so large that the question of what to accept, even tentatively,
is a serious one. The various estimates are assessed in the
following paragraphs, and suggestions are made about what to believe
and how to resolve the differences.

INDIVIDUAL COMPONENTS OF THE SURFACE HEAT BUDGET

Radiation Balance, F_r

Figure 18 graphically shows the disagreement in reported values of radiation balance and provides a basis for comparing them. The following features are revealing:

1. For the sunlit season, the radiation-balance values of Badgley, Chernigovskiy, and Marshunova agree almost exactly except for July, when Badgley's value is conspicuously lower. The reason is apparent if the values for absorbed solar radiation are compared. Both Soviet investigators report a lower albedo for July than for other months, with a consequent maximum of absorbed radiation in July rather than in June. The lower albedo of July is known to be real and should be taken into account. The United States program, however, has not included systematic observations of mean albedo during the melting season. Badgley's value for July is probably correct for an ice surface, but the Soviet values for July are probably more representative for the general area. Vowinckel's values are at the other extreme; he assumes a lower albedo and obtains much greater absorbed radiation.

2. The other conspicuous discrepancy is in winter values of net longwave radiation, where Badgley reports the smallest losses in fairly good agreement with Chernigovskiy. Marshunova and Vowinckel both report larger negative values. It is interesting that the former two authors base their values primarily on field observations at drifting stations while the latter two depend primarily on calculations. This suggests that such calculations either over-estimate the true surface temperature of the snow or underestimate the back radiation from the atmosphere during the coldest months.

The calculations are subject to uncertain knowledge about winter clouds and turbidity. Greater cloudiness or turbidity would cause values to be less negative. Chernigovskiy's values are based on data from seven Soviet drifting stations. Badgley's are based on one or two. In view of these considerations Chernigovskiy's
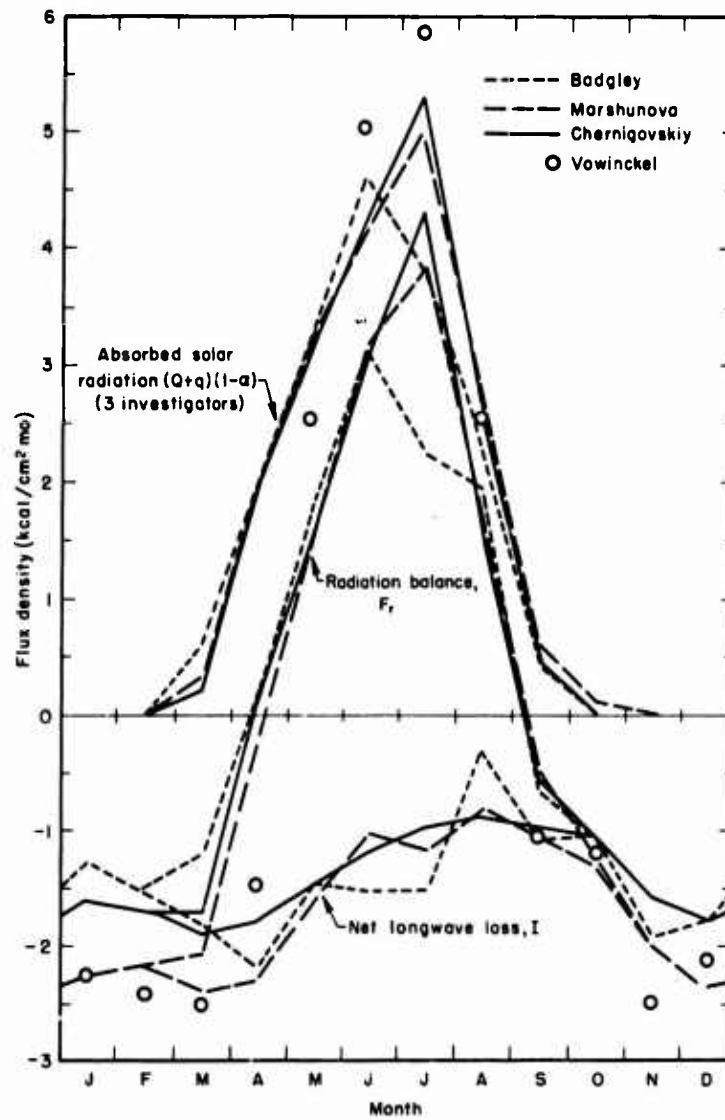
Fig. 18 -- Comparison of radiation-balance components
reported by various investigators.

Table 8

HEAT BUDGET OF ICE AND SNOW SURFACES* 82°--86°N

In this table (+) indicates energy added to the ice surface; (-) indicates energy leaving floe.

(average flux in kcal/cm^2 mo)

Type of Flux	Month												Annual
	J	F	M	A	M	J	J	A	S	O	N	D	
Net solar radiation (incident minus reflected)	0	0	+0.605	+1.99	+3.32	+4.63	+3.76	+2.24	+0.389	0	0	0	16.93
Net longwave radiation (incident minus emitted)	-1.30	-1.51	-1.81	-2.20	-1.47	-1.51	-1.51	-0.302	-1.08	-1.08	-1.94	-1.81	-17.52
Net radiation (visible plus longwave)	-1.30	-1.51	-1.21	-0.216	+1.86	+3.11	+2.24	+1.94	-0.692	-1.08	-1.94	-1.81	-0.06
Turbulent heat flux	+0.692	+0.562	+0.605	0	-0.562	-0.605	0	-0.173	-0.302	+0.0864	+0.259	+0.389	+0.95
Latent-heat flux	0	0	0	0	-0.432	-0.562	-0.432	-0.0432	+0.216	+0.259	+0.176	+0.0432	+0.775

* According to Badgley.

values seem the most likely, but the discrepancy between observed back radiation and calculated back radiation during winter should be further investigated.[*]

3. The unconnected circles in Fig. 18 are the radiation balance reported by Vowinckel for 90°N. They suggest that he greatly over-estimates the amount of solar radiation absorbed during summer. His values for net longwave radiation in winter are in good agreement with Marshunova's but slightly more negative.

4. These comparisons underscore the earlier remarks about the need for more field observations on:

       (a) albedo as a function of time and space during summer,

       (b) winter cloudiness (type, extent, height),

       (c) radiative properties of Arctic clouds.

Turbulent Heat Exchange Between Atmosphere and Surface, F_s

Since no one has yet measured the fluctuations of temperature and vertical air movement near the surface with the speed and precision necessary for a direct determination of turbulent heat flux, these values are obtained indirectly. The most common method is to measure the average gradients of wind speed near the surface to establish a coefficient of eddy diffusivity. The flux is then obtained as the product of this coefficient and the gradient of the transferable quantity (heat or moisture). This is the method used by Yakovlev and by Badgley.

A different approach was used by Laikhtman[32] in 1959 for an analysis of the results from North Pole 4 and North Pole 5. He developed an elaborate method of calculation that relates the heat fluxes to macro-meteorological parameters. His values are much larger than Badgley's and greater than Yakovlev's, partly because

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[*]At least three contributing factors are very uncertain: (1) the mean cloud conditions during winter; (2) the amount of ice-crystal turbidity, not reported as cloudiness, (3) the contribution of the ozone layer to back radiation during winter, when water vapor is almost nil.

the (absolute) values of radiation balance used in his calculations
are abnormally large. However, Laikhtman's values are quoted by
Gavrilova[4] in 1963 as representing the heat budget of the central
Arctic. It is also interesting to note that Laikhtman is a cc .hor
of the data represented in Fig. 11, which suggests that his earlier
estimates were revised downward only slightly by 1962.

In 1963 Doronin[33] calculated evaporation and turbulent heat
exchange using the same basic data from the Soviet drifting stations
but with different formulas. Since his results are physically
plausible and represent both a recent and carefully considered method
and also a reasonable compromise of the results of other investigators,
it is suggested that Doronin's values be tentatively accepted for the
purpose of constructing a composite heat budget. His monthly values
are given in Table 9.

Figure 19 compares results obtained by different authors, in-
cluding Vowinckel, who computed F_{ℓ} according to empirical formulas.
The large uncertainty is obvious; Laikhtman and Vowinckel obtained
their values by calculation from reported mean meteorological para-
meters. Badgley and Yakovlev base their results more on data from
drifting stations. The latter approach suffers from formidable
measurement difficulties and also from uncertainty about how repre-
sentative such observations are for the general area. Calculation
from meteorological parameters suffers from considerable uncertainty
about the physics of the boundary-layer processes over the pack
ice.

The following remarks can be made regarding the comparison:

1. Laikhtman's values can be expected to be too large because
he used abnormally large values for radiation balance in his
calculations.

2. Vowinckel's winter values seem too low since turbulent heat
flux during this season is known to be large enough to create deep
inversions in the atmosphere. (See Section II.)

3. Badgley's results agree quite well with Doronin's except
for July, August, and September. During these months of heterogeneous

Table 9

MONTHLY VALUES OF HEAT FLUX TO THE SURFACE (cal/cm^2)

BY TURBULENT HEAT EXCHANGE, F_s, AND EVAPORATION,

F_ℓ, ACCORDING TO DORONIN (cgs UNITS)

Term	Annual	J	F	M	A	M	J	J	A	S	O	N	D
F_s	2708	1184	758	721	292	-446	-386	-304	-402	-169	103	564	793
F_ℓ	-3204	2	-19	-32	-94	-464	-704	-638	-663	-386	-188	-9	-9

Computed according to the formulas:

$$F_s = \frac{c_p\, \rho k_1}{\ln \dfrac{x + k_1 z_2}{x + k_1 z_3}} (T_2 - T_3) \qquad F_\ell = 6.22(10^{-4}) \frac{L \rho k_1}{\ln \dfrac{X + k_1 z_2}{X + k_1 z_3}} (e_2 - e_3)$$

$$k_1 = \frac{0.16\, u}{\ln Z/Z_0} \qquad\qquad x = 0.18 \text{ cm}^2/\text{sec}$$

where:

T = temperature, °C e = vapor pressure, mb

c_p = specific heat of air at L = latent heat of cal/g
constant pressure, cal/g°C   evaporation of water,

ρ = air density, g/cm^3 u = wind speed, cm/sec

Z = height, cm Z_0 = roughness parameter
(value not given)

and the subscripts 1, 2, 3 refer to values at corresponding heights z_1, z_2, z_3.

surface conditions, the difficulties of obtaining field measurements that represent mean surface conditions are enormous. It will be seen later that Doronin's values give a more consistent picture of the total heat budget. Badgley's values are probably representative for conditions at the ice-flow stations but may not be representative of mean conditions over the central Arctic.

Latent-Heat Flux, F_ℓ

Figure 20 compares the same investigator's findings regarding heat flow by evaporation and condensation at the surface. Badgley and Yakovlev obtained their results in a manner similar to that for turbulent flux of sensible heat. Vowinckel used Sverdrup's empirical formula $E = K(e_w - e_a)V$, where V is wind speed, and e_w, e_a are the water-vapor pressures at the surface and in the air, respectively. Laikhtman and Doronin used more complicated relationships. The following comments seem relevant.

Agreement between Vowinckel and Doronin is quite good except for July. Vowinckel's minimum in July is caused by the assumed stability of the surface layer due to cooling from below. However, during July, surface ponds are most extensive. These shallow ponds absorb much solar radiation, show pronounced thermal stratification, and can be significantly warmer at the surface than the surrounding ice; over such ponds superadiabatic lapse rates can cause strong mixing. Vowinckel points out that raising of assumed surface temperature by only one or two degrees would eliminate his summer minimum.

In September and October, the surface is cooled by radiative loss, with stable air stratification and high humidity near the surface. Under these conditions evaporation is reduced, and condensation on the surface can occur. Badgley shows a significant rate of condensation instead of evaporation for September and October, a result consistent with such a picture.

The generally observed freeze-up in late August is evidence that the surface is losing heat at this time. Since observations

Fig. 20 -- Latent-heat flux, F_ℓ, at the surface according to various investigators.
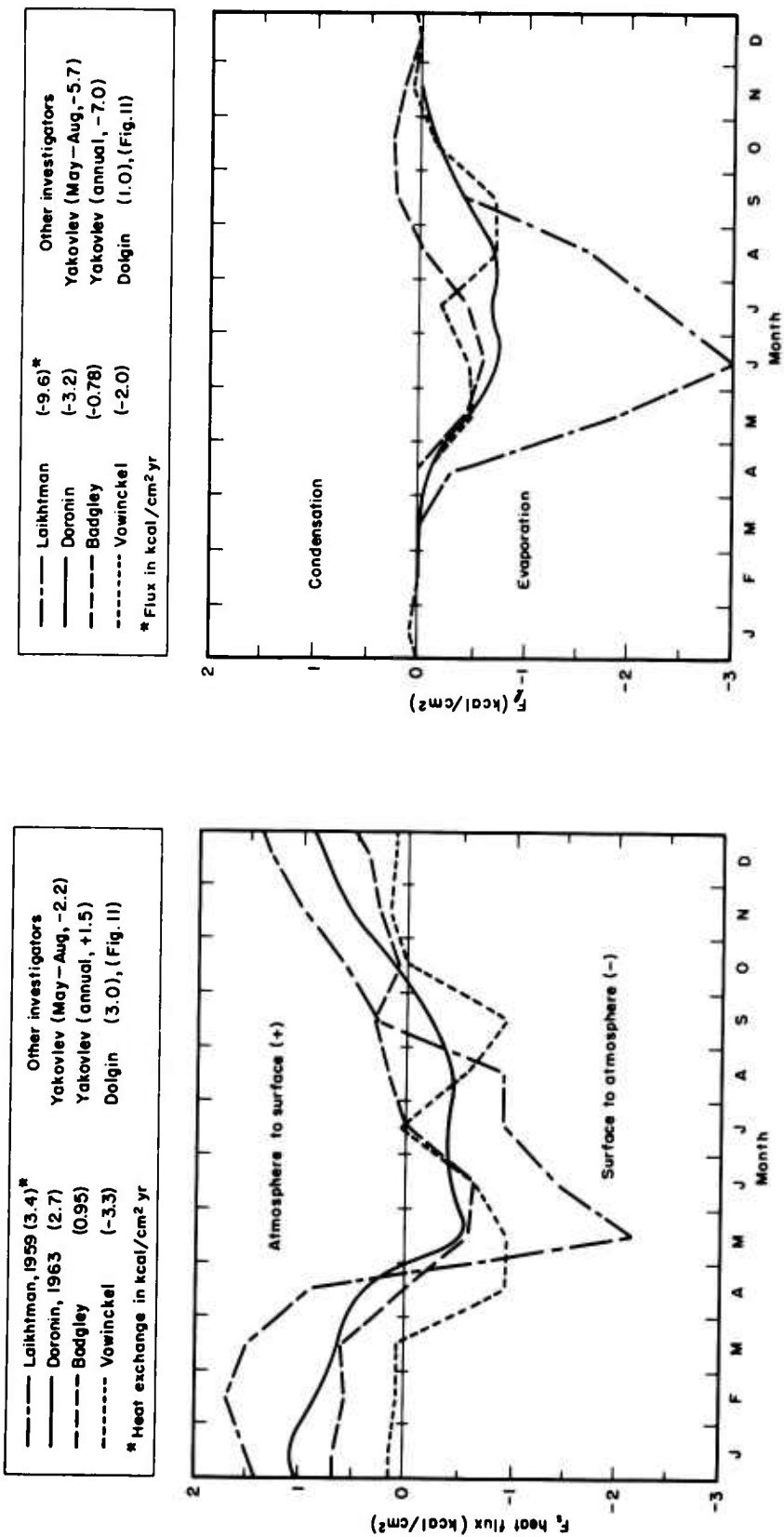
Fig. 19 -- Turbulent heat exchange, F_s, between the atmosphere and the surface according to various investigators.

show that the radiation balance is probably still positive at the end of August, there must be a significant heat loss by other processes. The rapid cooling of the atmosphere in August due to waning solar radiation should induce greater loss of sensible and latent heat at the surface, overcompensating the still positive radiation balance.

This evidence suggests that Badgley's values for evaporation in late summer are too small.

On the whole, Doronin's values provide the most consistent picture and are adopted for this study.

HEAT EXCHANGE BETWEEN THE SURFACE AND DEEPER LAYERS (MELTING, F_m, AND CONDUCTION, F_c)

The thermal regime of the pack ice is not so difficult to measure as is heat exchange with the atmosphere. Consequently it has been intensively studied and is better understood. Quantitative estimates can be viewed with more confidence than for the other components.

Beginning with North Pole 2,[34] almost every drifting station has measured vertical temperature distribution through the ice. The pattern observed at these stations during the course of the year is a consistent one and is closely approximated by the model described by Untersteiner.[28][29] This model applies to ice that has reached equilibrium thickness, with summer melt and winter freezing in balance. The model is mathematically complete, internally consistent, and describes observed features of the ice regime closely enough to be adopted for the purpose of this study.

According to this model, ice of equilibrium thickness adds 40 cm by freezing at the bottom each year and melts 30 cm at the top and 10 cm at the bottom. The heat available to the surface from below during the cooling period (20 August to 14 June) is shown in the accompanying tables.

In old ice:

Freezing of melt water retained on the surface	0.6 kcal/cm^2
Cooling of the ice mass	3.0 "
Heat of fusion from the bottom (40 cm)	2.7 "
Sensible heat from sea water	1.6 "
	7.9

In new ice and leads:

Heat released to the surface from freezing

of new ice in leads (30 cm)[*]	2.0 "
Total (kcal/cm^2)	9.9

[*]Although the percentage of open water is very small, the enormous heat loss gives this average for the surface as a whole.

The heat and ice quantities are balanced by:

Melting at the surface (30 cm of ice + 12 cm of snow + pond water)	3.7 kcal/cm^2
Melting at the bottom (10 cm)	0.7 "
Warming of the ice mass	3.0 "
Ice export by ocean currents (3000 km^3/yr = 30 cm)	2.0 "
Heat to atmosphere from old ice	0.5 "
Total (kcal/cm^2)	9.9

These heat transfers interact in a complex way (described by Untersteiner), so the actual passage of heat through the surface is of somewhat smaller amplitude. For example, some of the absorbed radiation passes through the surface, and the energy is distributed in the ice (1.2 kcal), and not all of the melting takes place at the surface. Untersteiner has calculated the actual heat flow at a level 10 cm below the surface (Fig. 20). His value of annual upward flow through this level is 5.3 kcal/cm^2. If we add the heat increments above this level (0.6 for freezing surface ponds and 0.1 for heat content of the upper 10 cm) we get 6.0 kcal/cm^2 rather than 7.9 reaching

the surface of old ice from below. However, to be consistent with our treatment of other components, we will assume that these heat transformations take place at the surface and will obtain a curve for heat exchange of the surface with deeper layers by adjusting Untersteiner's curve for the 10-cm level. Thus in Fig. 21 the total upward flow through the surface is 9.9 kcal/cm^2, and the downward flow is 6.3 kcal/cm^2. The difference between upward and downward heat flow (3.6 kcal/cm^2) comes from freezing of ice that is carried out of the Arctic Basin (2.0 kcal/cm^2) and from heat from the underlying ocean water (1.6 kcal/cm^2).

The shape of the curve for surface heat passage follows Untersteiner's curve through the winter months. For summer it is arbitrarily made linear, from zero in mid April to -2.75 in July to zero at the end of August. Such a pattern gives the right quantitative balance and corresponds to the generally observed maximum melting in July, freeze in late August, and ice warming from early April.

This curve for sensible-heat flux through the surface will be used to construct a composite heat balance for the central Arctic.

THE COMPOSITE HEAT BUDGET AT THE SURFACE

Having assessed each component separately, we can now consider a composite picture of the heat balance of the surface in the central Arctic. Figure 22 is such a representation and was constructed as follows:

1. The terms for ice melting and sensible heat exchange with deeper layers, $F_c + F_m$, for the pack ice as a whole are taken from Fig. 21.

2. The radiation balance, F_r, of Marshunova is taken from Fig. 13.

3. The remaining terms, F_s and F_ℓ, are obtained graphically in Fig. 22. They represent heat exchange between the atmosphere and the surface and are the most uncertain terms.
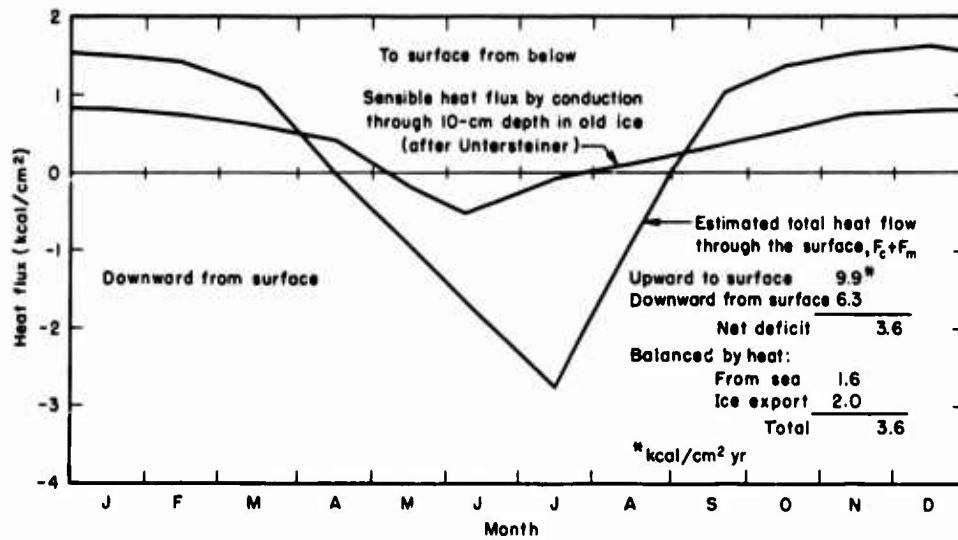
Fig. 21 -- Heat exchange between the surface and deeper layers.
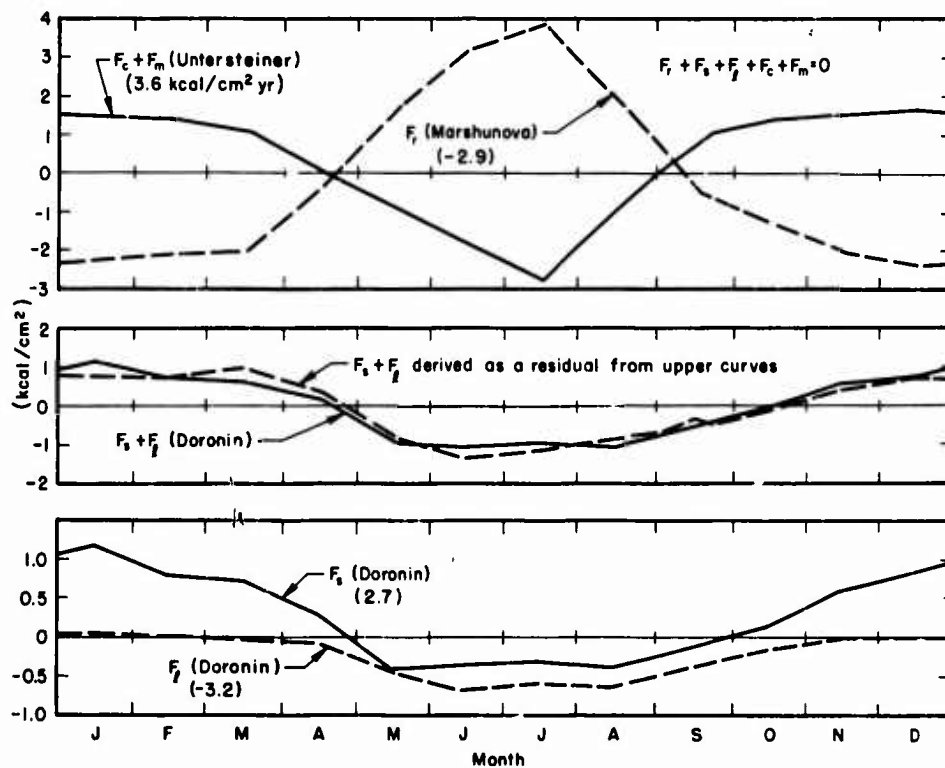
Fig. 22 -- The heat budget at the surface in the central Arctic.

4. In the middle section of Fig. 22, the values of $F_s + F_\ell$ reported by Doronin (Table 9) are compared with the values $F_s + F_\ell$ obtained as a residual. The agreement is quite good and can be taken as evidence that the composite heat budget represented by Fig. 22 is approximately correct.

CONCLUDING REMARKS ON THE HEAT BUDGET OF THE SURFACE

In several respects, the agreement shown in Fig. 22 is better than the basic data warrant. However, the pattern represented is internally consistent and can be taken as a reasonable description of the heat budget of the central Arctic.

It does not correspond closely with any one composite heat budget presented by other authors.

Except for the months of July, August, and September, Badgley's results are in generally good agreement. For those months his reported values are probably not representative of the central Arctic, although they may be representative of conditions over bare ice. As is noted earlier, his value of absorbed radiation for July is probably too small, as a result of assuming too high an albedo during the melting season. For the end of August he gives both a positive radiation balance (probably close to correct) and a flow of heat from the atmosphere to the surface. It is well verified that freeze-up occurs throughout the central Arctic during the latter half of August, so one of his values must be wrong.

Vowinckel's results differ drastically for radiation balance and are probably too positive in summer by a substantial margin. His values of latent-heat flux are in good agreement except for July, when his value seems too low. His values of sensible-heat flux seem much too small in winter, and in summer they appear too large except for July.

In Fig. 22 it should be noted that the annual radiation balance depicted is at the negative side of the probable true value (somewhere between -2 and +0.5 $kcal/cm^2$).

The discrepancy arises from Marshunova's values of net outgoing longwave radiation in winter (Fig. 18), which are more negative than Chernigovskiy's.

V. THE HEAT BUDGET OF AN ICE-FREE ARCTIC OCEAN

Several distinguished scientists have proposed that if the pack
ice were removed from the Arctic Ocean it would not immediately
re-form. [35--37] Instead, a different, and quasi-stable climatic
pattern would result. Such a process has been advanced as an explana-
tion for the Quaternary Period's ice ages, although theories vary
widely about the physical causes responsible for the disappearance of
the pack ice and for its later reappearance. Some causes suggested
have been variations in sea level (Ewing and Donn), obscuration of
solar heating by volcanic dust (Humphreys), and variations (for other
reasons) in incoming solar radiation (Simpson, Milankovitch).

In the past, so little was known about the Arctic heat budget
that these theories could not be assessed on any quantitative basis,
whereas qualitative arguments have been both inconclusive and unre-
vealing.

For these reasons a quantitative assessment of the heat budget
of an ice-free Arctic Ocean is of great practical interest both as a
test of alternative theories and as a basis for designing climatic
experiments (to be discussed later). As will be seen, current
estimates can vary widely. However, the author concludes that a
reasonable basis now exists for judging these estimates and that
further measurement and experimentation can substantially reduce the
uncertainties. The following paragraphs outline the issues involved.

THE HEAT BUDGET FOR AN OPEN LEAD

Badgley[28][29] has produced an estimate of the heat budget for
an open water surface at 82°--86°N under present climatic conditions.

Doronin[33] has also produced such an estimate. The values reported by these two authors are shown in Table 10.

The wide disagreement in values of latent and sensible heat fluxes indicates some of the difficulties attending such estimates and also the impropriety of taking values for an open lead as indicative of conditions for an open ocean. An open lead loses heat so rapidly in winter that it quickly freezes. The average amount of open water must not be larger than a small fraction of one per cent in winter; else the net heat loss would be too large to be compatible with what is known about the over-all heat budget.*

In Table 10, Badgley and Doronin agree closely on values for radiation balance, but their annual value is much less than the value we have used for an ice-free ocean. To see how this discrepancy comes about it is instructive to compare Badgley's basic data with values we have used. For solar radiation absorbed at the surface Badgley's value is 65; Fig. 7 shows 57; and Fig. 16 shows 58. This value is not very sensitive to area or latitude but is sensitive to cloudiness. Badgley uses present summer cloudiness in the central Arctic; Fig. 7 shows present summer cloudiness in the Barents Sea; and Fig. 16 shows an arbitrary 25 per cent depletion by clouds.

The values of net outgoing longwave radiation vary widely. Figures 7 and 16 show identical values of -37 $kcal/cm^2$ yr, but Badgley shows -72 in spite of the fact that his summer values are quite small. The wide discrepancy occurs because he has the relatively warm water radiating through a cold, dry, almost cloudless atmosphere in winter (typical of the present pack-ice climate) rather than to a moist, cloudy winter atmosphere. The result is that Badgley gets a net annual radiation balance of -7.25 $kcal/cm^2$ (with good agreement by

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*It is quite possible, however, that current estimates of ice export from the Arctic are much too low as a result of underestimating the mean thickness of the ice being exported by the East Greenland Current. When data is available from submarine transits better export estimates can be made. Since the heat budget of old ice must be approximately correct as given here, this increased export must be balanced by revised estimates of heat derived from freezing in open leads and thin ice.

Table 10

HEAT FLUXES FROM THE SURFACE OF AN OPEN LEAD ACCORDING
TO DORONIN (D)[33] AND BADGLEY (B)[28][29]

$(kcal/cm^2)$

Component	Source	Month					
		Jan	Feb	Mar	Apr	May	Jun
Solar radiation absorbed	B	0	0	2.2	8.9	16.5	15.8
Net long-wave loss	B	- 7.1	- 6.6	- 9.6	- 5.9	- 3.7	- 1.5
Radiation balance	D	- 9.5	- 8.5	- 8.2	0.2	9.6	14.4
	B	- 7.1	- 6.6	- 7.4	2.0	12.8	14.4
Sensible-heat flux	D	-168.0	-172.0	-177.4	-99.1	- 24.0	- 1.3
	B	- 19.4	- 17.3	- 13.4	- 8.6	- 4.3	- 0.6
Latent-heat flux	D	- 49.0	- 49.3	- 53.1	-37.2	- 15.1	- 1.9
	B	- 6.5	- 6.9	- 6.5	- 4.3	- 0.2	- 0.3

		Jul	Aug	Sep	Oct	Nov	Dec	Annual
Solar radiation absorbed	B	12.1	7.0	2.3	0.09	0	0	65
Net long-wave loss	B	- 1.0	- 1.2	- 5.8	- 6.9	- 9.3	- 12.6	- 72
Radiation balance	D	11.6	5.5	- 0.6	- 5.3	- 8.3	- 8.9	- 8
	B	11.1	5.7	- 3.5	- 6.8	- 9.3	- 12.6	- 7
Sensible-heat flux	D	- 0.2	- 1.4	- 11.7	-48.0	-120.0	-160.0	-983
	B	0	- 0.2	- 4.3	- 8.6	- 13.0	- 17.3	-107
Latent-heat flux	D	- 0.7	- 2.1	- 8.3	-23.4	- 41.9	- 51.2	333
	B	0	- 0.1	- 2.2	- 4.3	- 5.2	- 6.9	- 45

Doronin), whereas Fig. 7 shows +19.7 kcal/cm^2 and Fig. 16 shows 20.7 kcal/cm^2. This discrepancy not only illustrates the important influence that assumptions about winter cloudiness and humidity exert on estimates for an ice-free Arctic; it also indicates the probable necessity of eventually taking into account the associated changes in the atmospheric-circulation pattern. On the basis of present knowledge, however, we may conclude that a reasonable estimate for radiation balance for an ice-free Arctic with cloud conditions similar to those of the Barents Sea is about +20 kcal/cm^2 and that estimates based on open leads are very unrepresentative indeed.

Similar conclusions apply to values for sensible and latent heat fluxes to the atmosphere. Evaporation and turbulent heat exchange from the warm water to the cold, dry air reach huge values locally, but because the area of open water is small, the air, in the large, remains cold and dry. If the whole area were open water, such fluxes would quickly modify the air mass and suppress the heat loss by reducing the gradients of moisture and temperature near the surface. The enormous values given by Doronin also indicate the importance of assumptions about atmospheric stratification in such computations.

In short, rather than using the estimates for an open lead as representative of an open ocean, a more realistic basis must be found, one that takes into account the modification of the air masses as they pass over an extensive ocean surface.

As cold polar continental air moves over an ocean it is rapidly warmed and moistened from below, and these effects transform the initial inversion to an approximately adiabatic lapse rate up to the top of the layer of convective mixing.

The main questions are: How rapidly does this take place, and what are the mean conditions at the surface boundary as the air mass continues over the water surface?

MODIFICATION OF POLAR CONTINENTAL AIR OVER OCEANS

The theory of modification of polar continental air passing over oceans has been discussed by several authors.[38] Carefully selected

cases of air-mass modification over Hudson Bay have been analyzed by Burbidge, [39] who found that cold polar air masses crossing Hudson Bay in winter were modified during a 24-hour crossing to reduce the air--surface temperature difference by more than half. Craddock[40] analyzed trajectories of air masses from Iceland to Britain and found both heat and water-vapor flux to be proportional to the air--surface temperature difference. In the light of these data we may, as a first approximation, assume that the initially high values of sensible- and latent-heat flux would be reduced by half during the first 300 miles over water and by half again during the next 300 miles.

Since the Arctic Ocean is 1500--1800 miles across, a typical trajectory might have several 300-mile segments, each with half the surface-flux values of the preceding segment. In such a case the average value for the whole trajectory would be about a third that of the first 300-mile segment.

REPRESENTATIVE VALUES FOR THE FIRST 300 MILES

Extreme temperature contrasts can be found at high latitudes between cold polar air masses and the sea.

Evaporation and sensible-heat losses for various latitudes in the Norwegian and Barents Seas have been computed by Vowinckel and Taylor[30] (Figs. 23, 24). Their results show the characteristic shape of the curves of evaporation and sensible heat flux for polar continental air over water. The broad maximum for either flux from October to April is 3--5 $kcal/cm^2$ mo, and the summer minimum is negative. These are extreme values because in winter the air--water contrast is most extreme. The warm, narrow Norwegian Sea is subject to the most extreme contrast with polar continental air. The authors also point out that the values could be smaller by a third, depending on whose data for surface temperatures are used.

Budyko[41] gives somewhat higher values of evaporation and sensible-heat flux for polar continental air beginning its trajectory over the north Pacific (Fig. 25), an area of very great contrast between the cold outbreaks from Siberia and the warm water of the Kuroshio current.
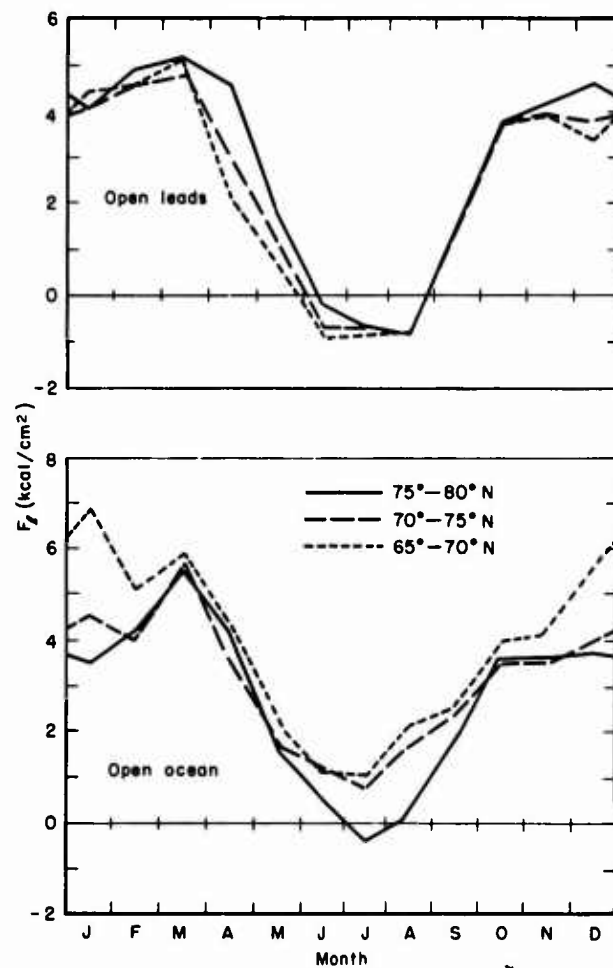
Fig. 23 -- Latent-heat flux in the Norwegian
          and Barents Seas (after Vowinckel).
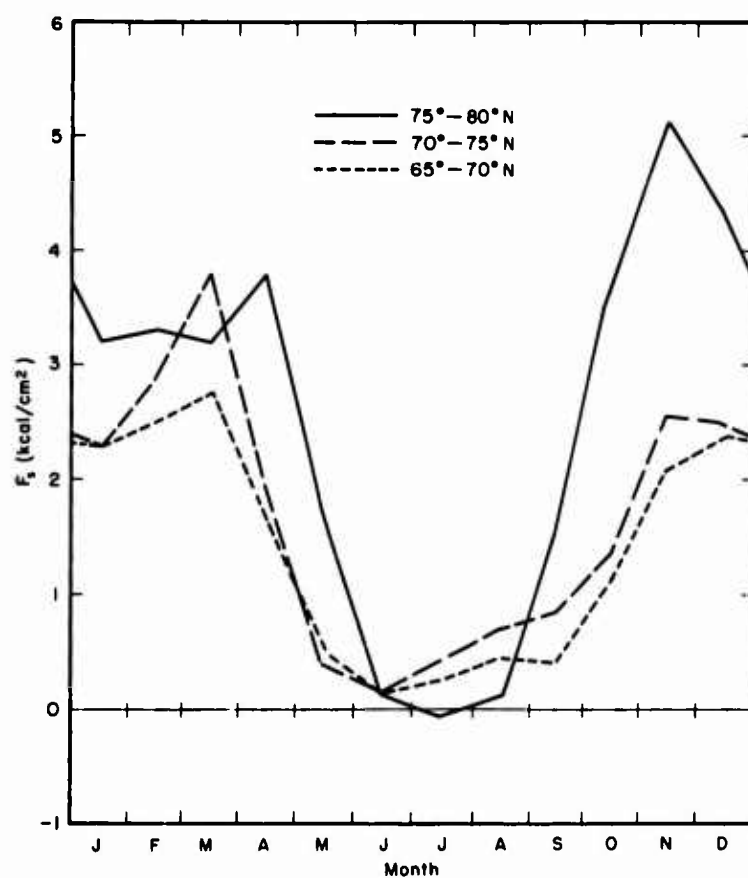
Fig. 24 -- Sensible-heat flux in the Norwegian
and Barents Seas (after Vowinckel).

ESTIMATED HEAT BUDGET FOR AN ICE-FREE ARCTIC OCEAN

Figure 26 is an estimate of the heat-balance components for the surface of an ice-free Arctic Ocean. It was constructed in the fol-lowing way.

(1) The radiation balance was taken from Fig. 16 and is probably the most reliable estimate in Fig. 26.

(2) Sensible- and latent-heat fluxes are taken as about a third of the mean values quoted from Budyko and from Vowinckel. These results seem reasonable when compared to values over oceans in Budyko's latest atlas.[12]

(3) To balance the heat budget at the surface, the residual is assumed to be the heat flux between the surface of the ocean and deeper levels, F_c.

In Fig. 26 the annual net heat exchange between the surface and deeper layers is about zero, suggesting that whether or not the pack ice would re-form depends on a fairly delicate balance (and may well depend on the advection of atmospheric heat). Assuming that the radiation balance, F_r, in Fig. 26 is approximately correct we may say that, if the winter heat losses from the surface to the atmosphere were much greater than shown, the pack ice would re-form. If they were much less than shown, the ocean would be gradually warmed until annual heat loss balanced annual heat gain.

We can say with confidence that present understanding of air--sea interaction is not reliable enough to indicate which way the balance would go. Further analysis of each of the components represented in Fig. 26 is needed.

BUDYKO'S METHOD OF ANALYSIS FOR AN ICE-FREE ARCTIC OCEAN

A quite different approach to the same problem has been taken by Professor M. I. Budyko.[*] He asks the question, "What would be the mean

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[*]Budyko is Director of the Main Geophysical Observatory, Leningrad. This institution, together with the Arctic and Antarctic Research Institute, Leningrad, has for many years placed major emphasis on studies of the Arctic heat budget, both with and without pack ice. It is of interest to consider their method and their conclusions.[42]
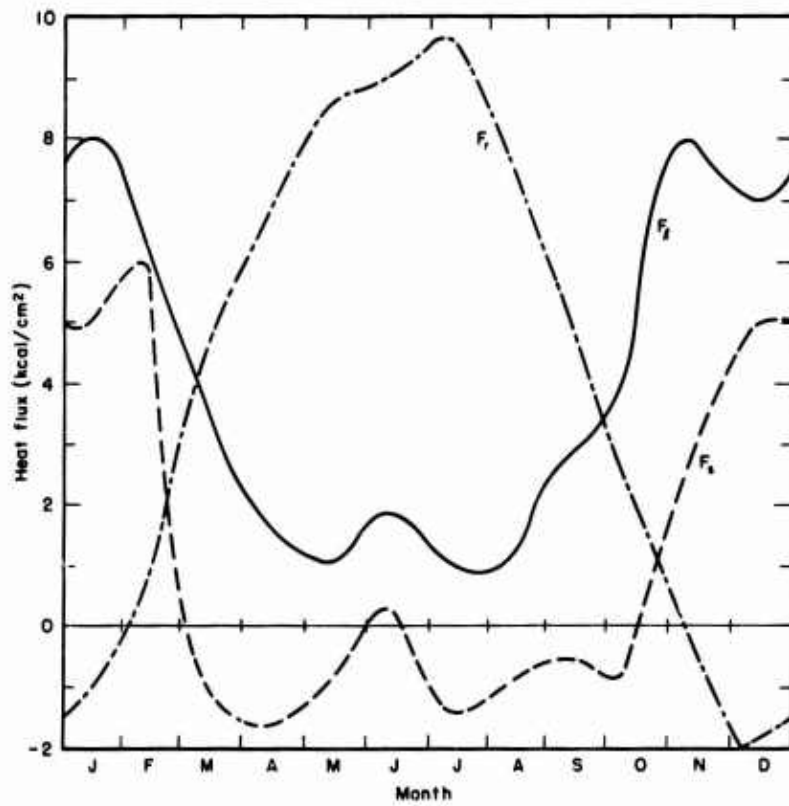
Fig. 25 -- Annual variation of latent- and sensible-heat
flux in the northwest Pacific (after Budyko).

air temperature near the surface for an ice-free Arctic Ocean?" He then tries to solve explicitly for air temperature by using the heat-balance method. His analysis begins with Eq. 5,

$$F_r + F_s + F_\ell + F_c + F_m = 0 \quad . \quad (5)$$

He then expresses the first three components in terms of meteorological variables as follows:

For fluxes of sensible and latent heat at the surface the widely accepted empirical relationships are of the form:

$$F_s = ac_p v(T_w - T) \quad ; \quad (6)$$

$$F_\ell = aLv(q_s - q) \quad , \quad (7)$$

where

a = coefficient of proportionality,

c_p = specific heat of air at constant pressure,

T_w, T = temperatures of the water surface and of the air, respectively,

v = wind speed,

L = latent heat of evaporation,

q = absolute humidity in the air, and

q_s = saturation humidity of air at the water-surface temperature T_w.

For the radiation balance at the surface, F_r, it is more difficult to introduce meteorological variables. This component is usually written as shown in Eq. (1):

$$F_r = (Q + q)(1 - \alpha) - I \quad , \quad (1)$$

where $(Q + q) = Q_0$ = total solar radiation at the surface,

α = surface albedo, and

I = net outgoing longwave radiation at the surface.
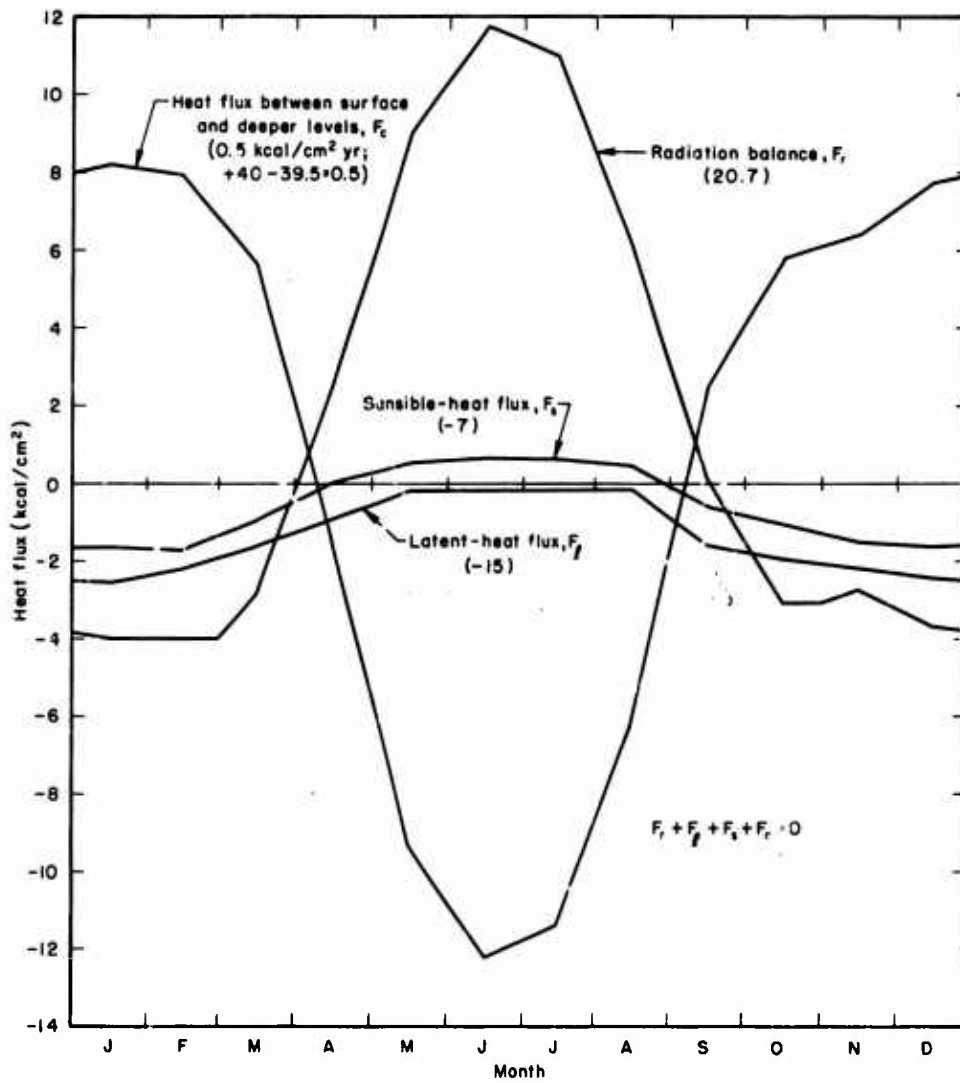
Fig. 26 -- Estimated heat budget components at the
surface for an ice-free Arctic Ocean.

The staff at the Main Geophysical Observatory conducted an elaborate investigation to find an empirical relationship to express the quantity I in terms of standard meteorological variables, using actinometric and meteorological data from the International Geophysical Year. They found that the following equation gives good results for computing monthly mean values:[43]

$$I = \epsilon\sigma T^4(11.7 - 0.40e)(1 - cn) + 4\epsilon\sigma T^3(T_w - T) \quad , \qquad (8)$$

where

ϵ = emissivity of the surface,

σ = Stefan--Boltzmann constant,

e = observed vapor pressure in mb,

c = empirical coefficient (numbers given by Marshunova), and

n = fraction of the sky covered by clouds.

Substituting these expressions for the corresponding components in Eq. (5) gives:

$$-(F_m + F_e) = Q_0(1 - \alpha) - \epsilon\sigma T^4(11.7 - 0.40e)(1 - cn)$$

$$- avL(q_s - q) - (avc_p + 4\epsilon\sigma T^3)(T_w - T) \quad . \qquad (9)$$

Equation (9) expresses the heat balance at the surface in terms of meteorological variables and turns out to be useful for assessing the interaction of air temperature and surface conditions.

For example, the amount of summer melting can be calculated as a function of air temperature. We will use this relationship later to estimate the effects of anomalous summer air advection on ice thickness.

When applied to the melting season Eq. (9) can be simplified by the following approximations:

1. The temperature difference between che surface and the air is small, relative humidity is high and almost constant. Thus we may write $q_s - q = \beta q_s \approx 0.1q_s$;

2. Surface temperature is almost constant and is near the melting temperature of ice, so $T_w \approx 0°C$;

3. The quantity F_c is small during the melting season (Fig. 21) and can be approximated.

Budyko used Eq. (9) to estimate the mean air temperature, T, corresponding to an ice-free Arctic Ocean. He wrote Eq. (9) in the form:

$$Q(1 - \alpha) - I_0 = avLq_s\beta + (avc_p + 4s\sigma T^3)(T_w - T) - (F_m + F_c) \quad ,$$

where

(10)

$$I_0 = \epsilon\sigma T^4(11.7 - 0.40e)(1 - cn) \quad .$$

He then deals with the question of how air masses are transformed in moving over the ocean by introducing a relationship developed by Berlyand,[44]

$$T'_1 - T = \gamma(T_w - T) \quad , \tag{11}$$

where

   T'_1 = average temperature of air passing through the periphery of the Arctic Ocean into the central region,

   T''_1 = same for air passing south through the periphery out of the central region,

   γ = coefficient of proportionality related to the dimensions of the water body and the speed of movement of the air masses.[*]

Budyko then makes the following approximations:

(1) The average air temperature at the periphery of the Arctic Ocean, T_1, is about equal to $(T'_1 + T''_1)/2$;

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[*]It may be noted that in another source Budyko[45] gives the value of 0.85--0.90 for this coefficient for the Arctic Ocean. That is to say, initial temperature difference between air and surface would be reduced 85--90 per cent on the average, rather than by two thirds as assumed in constructing Fig. 25. In such case the annual heat balance would be positive, and warming would take place.

(2) $T''_1 = T$.

Then,

$$T_1 = (T'_1 + T)/2 \quad . \tag{12}$$

Using Eqs. (10)--(12) Budyko calculated the air temperature in the central Arctic under present conditions. The value T_1 was taken to be equal to the mean annual air temperature at $70^O N$. The components of the radiation regime were taken as the mean of the values given by Marshunova for $80^O N$ (see Appendix); calculations were made for each month.

The mean annual air temperature calculated by this method turned out to be near the value observed for $80^O N$, thus lending some credence to the method.

The method was then applied to the case of an ice-free Arctic Ocean.[*] In this case the heat exchange between the surface and deeper layers can reach large values, whereas with an ice cover it is small. We can consider F_c as the sum of two components: the variation in heat storage in the water column, D_1, and an advective component brought in by ocean currents, D_2.

The first component, D_1, is equal to zero for the whole year and for individual months can be approximated by

$$D_1 = k(\overline{T}_w = T_w) \text{ kcal/cm}^2 \text{ mo} \quad ,$$

where $\overline{T}_w$ = mean annual water-surface temperature, and k = coefficient which equals, on the average, 3.1 kcal/cm^2 mo deg. The second component, D_2, was taken to be proportional to the meridional gradient of water temperature. The temperature of air masses passing northward into the Arctic under the no-ice condition was taken as the same as under present conditions.

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[*]Equations (11) and (12) give good results for an ice-covered sea, as claimed; there is some reason to doubt their validity for an open sea. As noted earlier, the pack ice is a barrier to heat exchange between the atmosphere and the sea. Comparing Fig. 22 with Fig. 26 suggests that transformation of air masses would proceed quite differently in the two cases.

Budyko's Results

As calculated by Budyko, the following conditions are to be expected of an ice-free Arctic.

1. In the central Arctic, air temperature in winter would be about $+5^{\circ}$C and in summer about $+10^{\circ}$C. Water temperature in the coldest month would be about $+5^{\circ}$C.

2. In the peripheral zone, air temperature in winter would be about -10°C and in summer $+10$ to $+15^{\circ}$C. Along shore a band of ice would form in winter to a width of several hundred kilometers. This ice would retard heat loss in winter but would be expected to disappear in May and hence affect the summer heat income very little.

Budyko claims that these conditions correspond well with what is known about the climate of the upper Tertiary, just preceding the onset of the Quaternary's ice ages.

Using the method described above, he maintains that his calculations show that a relatively small initial lowering of air temperature in times before the Arctic was ice-covered would cause substantial widening and thickening of the ice bands along the shore, thus delaying the melting in summer and lowering the mean albedo during the critical period of solar heating. This led to a further reduction of temperature and even more development of the shore ice -- until a large part of the Arctic Basin was ice-covered. Thus the ice-free condition was relatively unstable in the sense that when the critical heat budget was approached, conditions could change quickly.

Budyko's results and conclusions contain some rather large assumptions. They do, however, emphasize the need for further efforts to assess the individual components of Fig. 26, which in its present form can serve as a first approximation to the heat budget of an ice-free Arctic Ocean.

VI. THE ARCTIC HEAT BUDGET AND THE CIRCULATION
OF THE ATMOSPHERE

One objective of this study is to explore the question: "Does
the Arctic heat budget offer possibilities for either predicting or
influencing climatic conditions?" Up to this point, we have discussed
only the heat budget. Our knowledge is still sparse and some of the
quantitative estimates are uncertain. Much work needs to be done to
reveal the physical processes at work and to reduce the quantitative
uncertainties. The author hopes that calling attention to specific
needs and the context in which they appear will encourage interest
in these efforts.

We do know enough about the Arctic heat budget to consider the
interaction of its components with the circulation of the atmosphere,
and with its climate-forming influences at lower latitudes. Later,
we will discuss an experimental approach to these questions.

In this section, we will lay the groundwork for such an approach
by considering the gross aspects of the two questions:

(1) What is the influence of the pack ice on the atmospheric
circulation and the climate of lower latitudes?

(2) What is the influence of atmospheric circulation on the
thickness and extent of the pack ice?

DOES THE PRESENCE OF THE PACK ICE CAUSE
CIRCULATION INTENSITY TO INCREASE OR DECREASE?

The usual discussion of the role of the pack ice in climatic
change points out that the high albedo of the snow-covered pack de-
creases solar heating, increases the regional heat deficit, and by
cooling the atmosphere, intensifies meridional temperature gradients

and zonal circulation. Thus, more extensive ice contributes to stronger atmospheric circulation.

Usually no distinction is made between summer and winter, but since solar heating is the issue, the effect should be most evident in summer.

This perfectly logical line of reasoning leads to conclusions that are probably almost opposite to the true behavior of nature. The following paragraphs explain why atmospheric circulation intensity should be greater rather than less if the pack ice were not present. This surprising conclusion depends on the quantitative assessment of the annual pattern of atmospheric heat loss.

THE PACK ICE AS A HEAT BARRIER BETWEEN OCEAN AND ATMOSPHERE

Comparing Fig. 26 with Fig. 22 shows how effectively the pack ice reduces the heat exchange through the surface by radiative and nonradiative processes. From Fig. 26 we see that with no ice, about 40 kcal/cm^2 is stored in the ocean during summer and is released at the surface during winter. From the table preceding Fig. 21 we see that, with pack ice present, only about 9.9 kcal/cm^2 is released at the surface in winter and replenished in summer, and only 1.6 kcal/cm^2 of this comes from ocean waters.

The ice itself accounts for the rest of the heat storage; by cooling (3 kcal/cm^2) and freezing (5.3 kcal/cm^2). Thus, while the ice sheet undergoes large seasonal changes in thermal state, the ocean waters are in relative thermal isolation, slowly losing heat upward during almost the whole year. The total heat passage in both directions at the top of the water column (1.6 kcal/cm^2) is only about a fiftieth as much as with no ice (80 kcal/cm^2 for both directions). Because of this, the presence of the pack ice exerts great influence on the structure and the dynamics of both the ocean and the atmosphere in the Arctic.

THE ANNUAL PATTERNS OF ATMOSPHERIC HEAT LOSS

During the long winter darkness, the pack ice limits the heat flow from the ocean, so the heat loss to space must come mainly from the atmosphere. This atmospheric cooling increases the temperature gradients between the Arctic and lower latitudes, thus causing the atmospheric circulation in winter to be more intense and bringing more heat to the Arctic by advection.[*] In this way, the pattern of atmospheric cooling in the Arctic influences the general atmospheric circulation.

It is instructive to compare the annual pattern of atmospheric heat loss with and without pack ice. The value of the loss of atmospheric heat can be readily obtained by subtracting the heat supplied to the surface from below from the heat balance of the earth--atmosphere system. The heat fluxes from below are taken from Fig. 21 and Fig. 26. For the heat balance of the earth--atmosphere system corresponding to ice-covered and ice-free oceans, the values shown in Fig. 10 and Fig. 17 are used.

The results of these computations are shown in the upper portion of Fig. 27. In the lower portion are shown the annual variation of several indices of atmospheric circulation under present (ice-in) conditions.

THE PATTERN OF HEAT LOSS UNDER PRESENT (ICE-IN) CONDITIONS

With pack ice, relatively little heat from the ocean escapes to the atmosphere, so the losses to space are borne almost entirely by the Arctic atmosphere itself. In summer, the pack ice inhibits storage of solar heat in the ocean. Considerable solar heat is absorbed by the atmosphere; some, transformed at the surface, enters the atmosphere as sensible and latent heat. This warming in the Arctic reduces temperature gradients between the Arctic and peripheral areas.

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[*]A heat loss of 240 cal/cm^2 cools the whole atmospheric column by about 1°C.

The pattern representing the ice-in condition is typical not only of the pack-ice area but of any high-latitude area in which heat flow from below the surface is small. Thus under present climatic conditions loss of atmospheric heat from the whole snow-covered area north of, say 60°, is represented roughly by the ice-in curve.[*] As shown by the lower curves in Fig. 27, the intensity of atmospheric circulation follows a similar course, with intensification of both zonal and meridional circulation in winter at mid latitudes and corresponding weakening in summer.

Direct calculations of atmospheric heat advection to high latitudes are laborious and, therefore, scarce, although the radiosonde coverage is fairly good around the periphery of the Arctic Ocean. However, such values as have been reported are in good agreement with the derived curve for atmospheric heat loss in Fig. 27.

For transport of sensible heat across 70^ON, Bjerknes and Mintz reported 8 kcal/cm^2 mo for January--February, 1949, and 1.2 kcal/cm^2 mo for July--August, 1949.[47] These values are shown on Fig. 27.

For transport of both sensible and latent heat across 70^ON, Starr and White reported 81 kcal/cm^2 for the entire year 1950,[48] and Houghton has reported 78 kcal/cm^2 as an annual value.[14] It can be seen from Fig. 27 that these values are in quite good agreement with our computed losses of atmospheric heat.

THE PATTERN OF ATMOSPHERIC HEAT LOSS
FOR AN ICE-FREE ARCTIC

The annual patterns of atmospheric cooling for ice-in and ice-out conditions are different in several important ways.

The most striking difference is the relatively high rate of atmospheric cooling in summer over an ice-free Arctic Ocean. Thus meridional temperature gradients should be greater rather than less

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[*]In the Norwegian Sea the ice boundary is far to the north, but in the continental areas it extends far south of 60^ON. On the whole, the snow-covered area is larger than the area north of 60°.
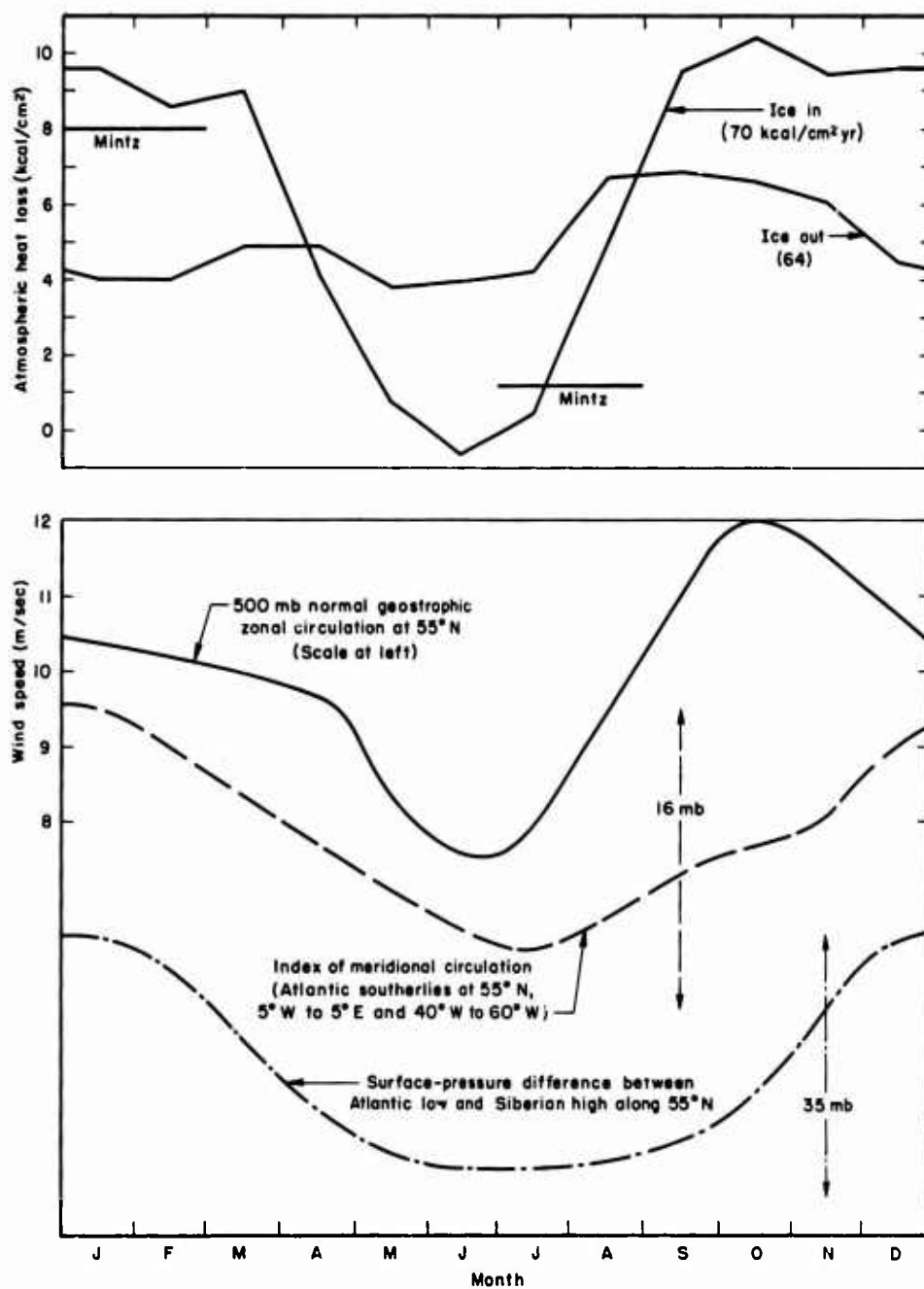
Fig. 27 -- Monthly requirement of atmospheric heat for the central
Arctic (above) and indices of atmospheric circulation
(below). Indices taken from U.S. Weather Bureau's
normal weather charts.

and atmospheric circulation stronger rather than weaker. The reason
for this seeming paradox is that while income of solar radiation is
high for the earth--atmosphere system, without pack ice most of this
heat goes into the ocean -- to be released to the atmosphere during
winter.

On the other hand, in winter an open Arctic Ocean furnishes much
of the heat loss of the earth--atmosphere system and reduces atmos-
pheric cooling to about half of the winter value for an ice-covered
Arctic Ocean. Thus in winter atmospheric circulation should be
weaker -- but how much weaker? We must remember that the temperature
difference between air entering and leaving the region would be much
less, for over an open sea the lower three or four kilometers should
be near moist adiabatic conditions with a surface temperature at
freezing or above. As a rough approximation it seems the difference
in heat content between air entering and leaving the region might be
reduced by half. If so, even half as much atmospheric heat loss
would require just as much air mass exchange as under present con-
ditions. Thus the reduced atmospheric heat loss in winter over an
open Arctic could not correspond to much decrease in circulation
intensity.*

THE GENERAL CLIMATE OF AN ICE-FREE ARCTIC

If we consider the year as a whole we note that the difference
in atmospheric heat loss between ice in and ice out is only 6 kcal/cm^2,
a reduction of about 10 per cent from present conditions. Of far
greater importance would be the change in the annual pattern of
atmospheric cooling, with the more vigorous atmospheric circulation
in summer more than offsetting a less vigorous winter circulation.

What is most striking about the pattern of heat loss for an ice-
free Arctic is the relative constancy during the year, implying a

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*At this point the reader may ask, "What then, drives the cir-
culation? How would the atmosphere adjust to such conditions?" An
experimental approach to these questions will be discussed in Section
VIII. For the moment it may be pointed out that a moist, cloudy
atmosphere cools at high levels, a dry clear atmosphere at low levels
(present conditions). Thus meridional gradients above three or four
kilometers might be even greater than at present.
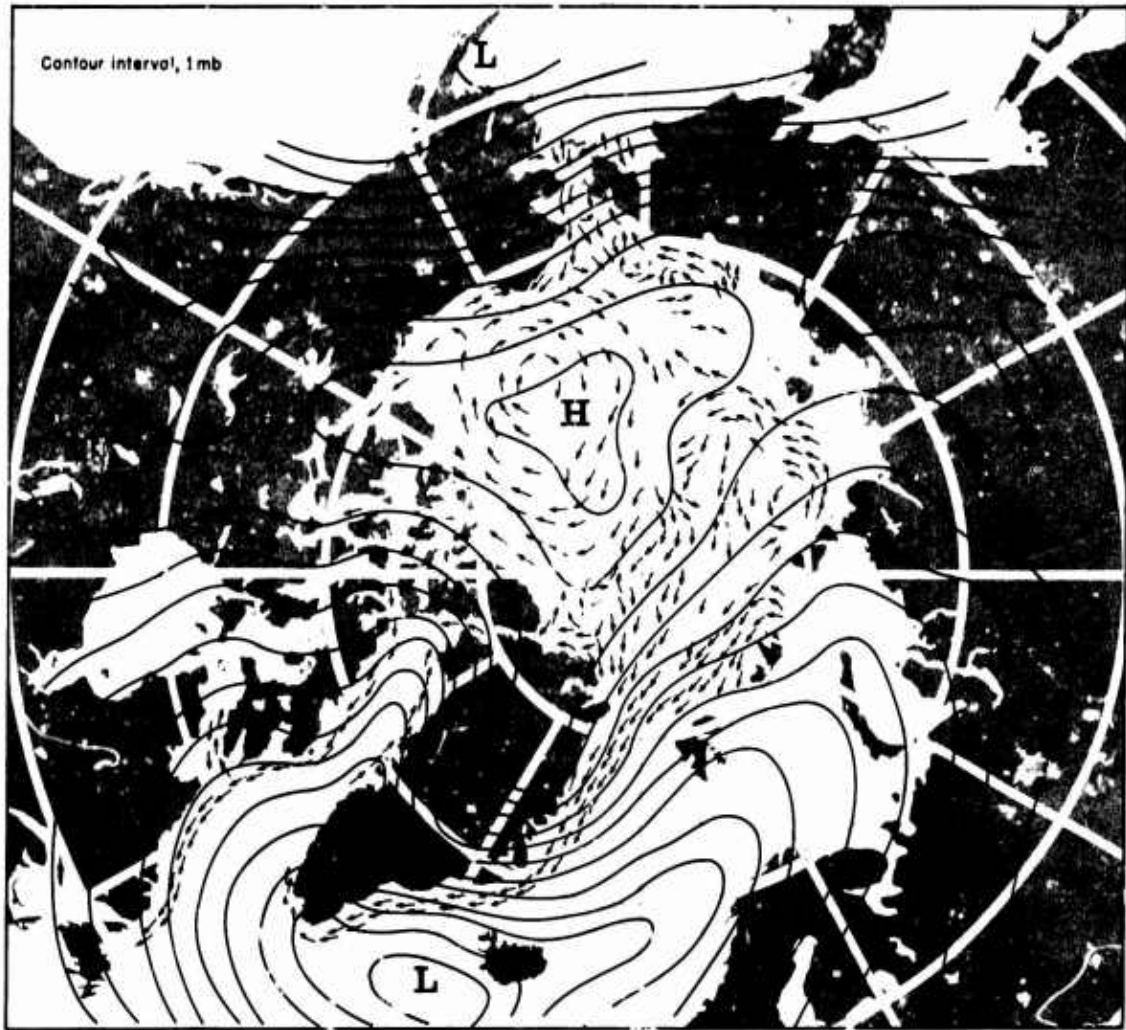
Fig. 28 -- Annual mean patterns of surface pressure and ice movement (after Dunbar and Wittman[49]).

more monsoonal type of circulation throughout the year. At mid latitudes summers should be cooler and winters warmer than at present.[*]

The combination of relatively vigorous and more constant zonal circulation with cool moist summers and warm moist winters would be highly favorable conditions for accumulating continental ice sheets at high latitudes.

With pack ice present, evaporation is small and occurs mostly in summer, so snowfall from Arctic air masses is very light indeed. With an ice-free Arctic Ocean, the cold polar continental air in winter would be rapidly transformed to polar maritime air, moisture laden and with unstable lapse rates in the lower third to half of the atmosphere. Almost continuous snowfall could be expected along coastlines with on-shore winds, conditions generally similar to the Great Lakes Snow Belt, the Great Snow Belt of Japan, or the eastern shore of Hudson Bay.

The total amount of snowfall can be estimated from Fig. 26, which shows mean evaporation from the whole Arctic Ocean of about 30 cm, almost all of which is in winter.[**] This is equivalent to about 3 m of fresh snow over an equal area (or 9 m over an area a third as large). Accumulation of such quantities at higher elevation, where summer melting would be reduced, could lead to the rapid development of continental ice sheets. The crucial questions are what the atmospheric circulation pattern would be and where heavy snowfall would occur; these are matters for experimental investigation, but some qualitative speculation at this point may be useful in designing such experiments.

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[*]Since conditions leading to the present low-level outbreaks of cold polar air would not be present, the warmer mid-latitude winters might well correspond merely to the absence of such outbreaks.

[**]For inputs to general circulation experiments it must be remembered that most of this evaporation and surface heating would occur near shore.

REMARKS ON THE PRESENT PATTERN OF ATMOSPHERIC CIRCULATION

In the larger sense, climate is determined by three basic factors: influx of solar heat, circulation of the atmosphere, and configuration of continents and oceans. Atmospheric circulation, in turn, is controlled by irregular cooling and heating, rotation of the earth, and configuration of continents and oceans. In the Northern Hemisphere, two meridionally extended continents with high mountain barriers and two oceans influence the location of the main atmospheric centers of action; the Bermuda high and Icelandic low in the Atlantic, the Pacific high and Aleutian low in the Pacific.

Figure 28 shows the mean annual surface pattern of atmospheric and oceanic circulation at high latitudes. It should be remembered that in such a time-averaged mean, the troughs correspond roughly to cyclone paths; the low centers are areas of cyclogenesis, with lows deepening and slowing as they pass.

The Icelandic and Aleutian lows are at about the same latitude and are similarly situated with regard to the thermal contrast between ocean and land masses to their northwest. However, this contrast is greatly weakened with summer heating of Siberia, causing the Aleutian low in summer to become much weaker than the Icelandic low, which is flanked by the cold, high plateau of Greenland. Thus in the annual mean, the Icelandic low is especially conspicuous. This anchoring of the Icelandic low in both space and time by Greenland is important when possible changes in atmospheric circulation are being considered.

Figure 29 shows the mid-winter pattern in the Atlantic sector and Fig. 30 the mid-winter pattern in the Pacific sector.

The main winter cyclone tracks are shown in Fig. 31 The boundary between the pack ice and the relatively warm and moist Norwegian and Barents Seas is the avenue by which most cyclones migrate northward. Some of these cyclones originate in the southeast United States and some as far away as Japan, coming by way of the Aleutian trough and across central Canada. A few cyclones enter the Arctic Basin via Bering Strait to join the giant spiral that ends in northern Canada. Also cyclones often move northward west of Greenland where they usually die.

The summer pattern (Fig. 32) is similar except that the main cyclone tracks tend to be more zonal in northern Europe, following the thermal contrast between the cold Arctic Basin and the warm continent.

Thus the eventual destination of most cyclones is the northern Canadian islands. In spite of this pattern, present winter snowfall in this region is very light because the low winter temperatures reduce the moisture associated with these cyclones long before they reach this area.

ATMOSPHERIC CIRCULATION ASSOCIATED WITH
AN ICE-FREE ARCTIC

Although the thermal regime of an ice-free Arctic would be markedly different from present conditions, we should expect the main features of the hemispheric circulation to remain much the same because their physical causes would remain as before. The Aleutian and Icelandic cyclone tracks seem to be closely related to the thermal contrasts off the eastern margins of the continents and to the positions of the great mountain barriers, all factors that would be essentially unchanged.

The most obvious difference to be expected in the mean pressure pattern is the elimination of the winter high-pressure area over the Arctic Basin (Fig. 28). The cold winter highs over the continental areas should remain, but the relatively warm Arctic Basin should be a region of cyclonic activity. The cyclones from the north Pacific could move more easily into the Arctic instead of being deflected to the south. The cyclones moving up the Norwegian Sea would no longer be "dried out" upon entering the cold Arctic (in the Barents and Kara Seas) but would continue across the Arctic. They would be blocked by continental highs over Siberia and Alaska, but between Alaska and Greenland, instead of a semipermanent ridge as at present, there should be a permanent trough separating the Greenland anticyclone from the Alaskan ridge. This trough west of Greenland should be subject to especially heavy snowfall, for it would be the natural destination of both the Atlantic and the Pacific cyclones.
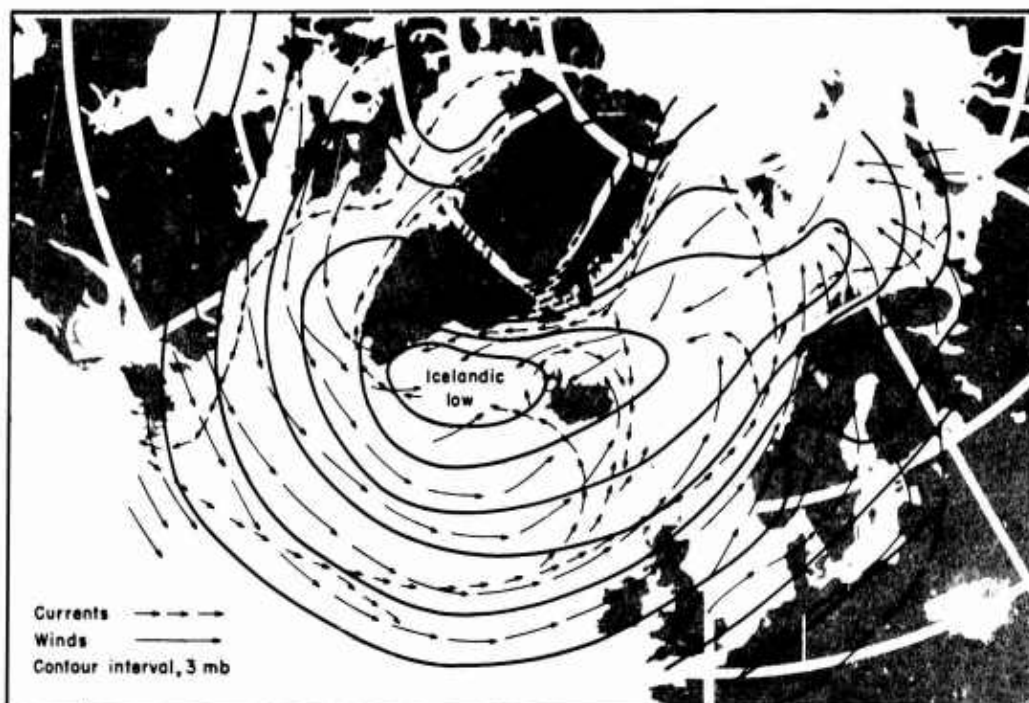
Fig. 29 -- Schematic pattern of ocean currents in winter months and prevailing surface winds in January (North Atlantic).
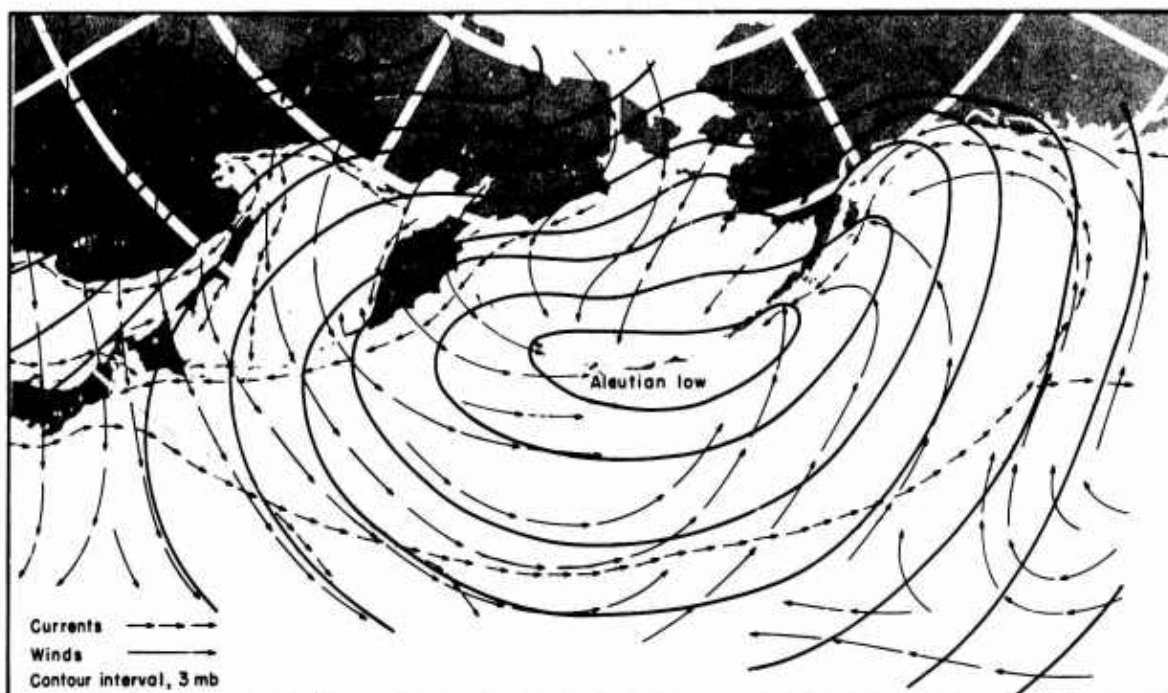
Fig. 30 -- Schematic pattern of ocean currents in winter months and prevailing surface winds in January (North Pacific).

POSSIBLE ACCUMULATION OF CONTINENTAL ICE SHEETS

It should be noted that the 8000-ft elevations of Ellesmere
Island are unique in northern Canada, and the higher elevations
extend southward toward eastern Baffin Island. Under the comparatively
mild climate of an open Arctic Sea, much more snow can be expected
to persist through the summer at high elevations than at low levels.
The combination of high elevation and very heavy winter snowfall
would be most conducive to glacial accumulation. The great conti-
nental ice sheets could well have originated in this way.*

*Study of the climatic history of Ellesmere Island should be most
revealing, especially when correlated with ocean-bottom cores from the
Arctic Ocean and the North Atlantic. For the results of preliminary
studies see Müller.[51]

Fig. 31 -- Principal tracks of low-pressure areas
in January (after Wilson[50]).
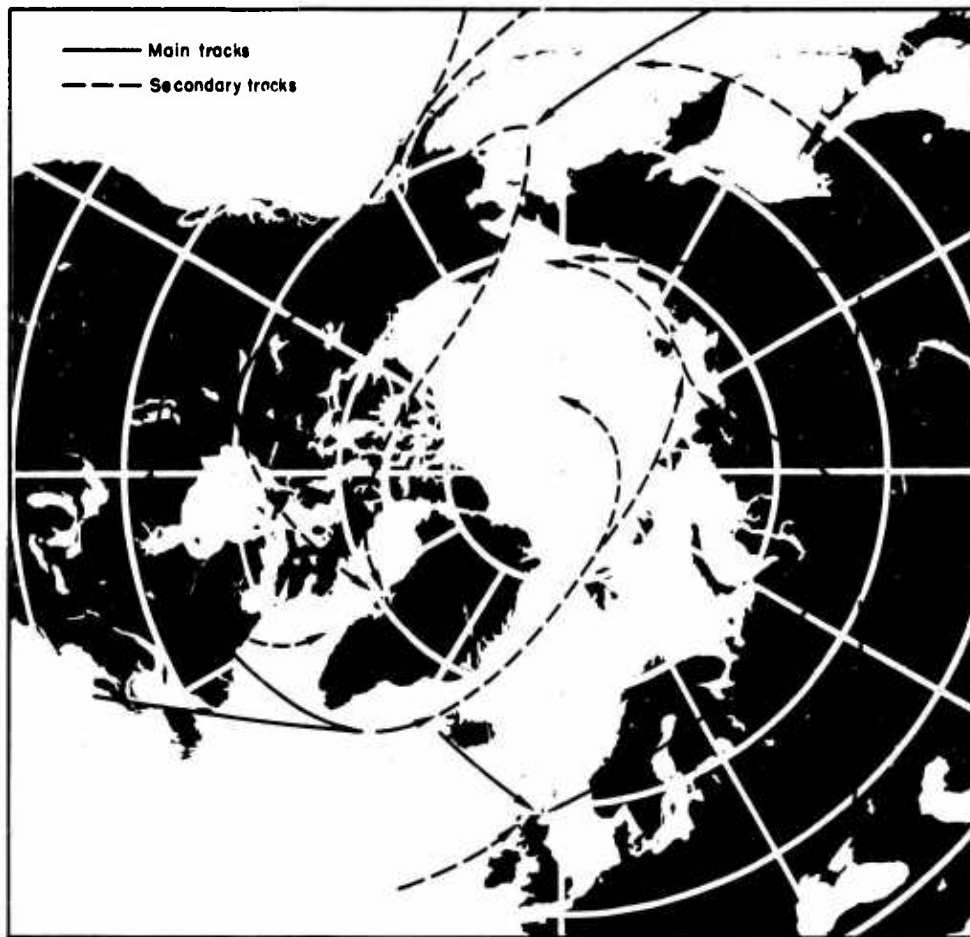
Fig. 32 -- Principal tracks of low-pressure areas
in July (after Wilson[50]).

According to Fig. 26, evaporation from the open Arctic Ocean would occur almost entirely in winter and would yield about 3 m of snow over an area comparable to the Arctic Ocean. Some fraction of this would be deposited on land. In addition, moisture brought in by the Atlantic cyclones could be expected to cross the Arctic instead of being dried out by low temperatures as at present. Thus very heavy snowfall could be expected along the Quebec--Ellesmere meridian, building outward towards the west.

Possible Termination of an Open-Arctic Period

The Pleistocene ice, at its maximum extent, covered almost all of Canada and a somewhat smaller area of northwest Eurasia, with a general advance of mountain glaciers all over the earth; according to Flint,[52] 45 million square kilometers of the earth's surface were covered with the ice, as compared to 15 million today. Almost all of this additional 30 million square kilometers was in the Northern Hemisphere, about 12 per cent of the hemisphere. If we assume an albedo change for the earth--atmosphere system from 0.4 (over forest or tundra) to 0.7 (over dry glacier), it would require a drop of about $3.5^{\circ}C$ in the mean radiating temperature of the hemisphere to maintain radiative equilibrium. By this rough estimate it is plain that such extensive ice fields would greatly alter the radiative balance of the earth.

Since the continental ice sheets almost surround the Arctic Basin, their cooling effect should be concentrated there. Such a process seems a more probable cause of refreezing than decreased oceanic advection due to lowering sea levels, as suggested by Ewing and Donn. One reason is that the additional cooling of the ice sheets would be a heat loss about two orders of magnitude greater than the total oceanic heat transport into the Arctic. Other reasons are discussed in Section VII.

The relatively small drop in temperature that Budyko claims would cause the pack ice to form could easily be caused by the lower albedo of extensive continental ice sheets.

INFLUENCE OF THE ATMOSPHERE ON THE PACK-ICE REGIME

We now come to the second question posed at the beginning of this section: what is the influence of atmospheric circulation on the thickness and extent of the pack ice?

Laikhtman has investigated the effects of air-mass advection on the speed of ice melting. [32] He found that the main factors controlling the amount of melting are radiation balance and air temperature. During changes in humidity and wind speed in the air masses moving over the Arctic, there is a mutual compensation of the rate of evaporation and sensible-heat exchange, so variations in humidity and wind speed have relatively small influence on the net heat exchange at the surface. Especially during summer it is the tempera- ture of advected air that is most influential in causing changes in ice thickness.

The quantitative influence of atmospheric advection on the ice regime can be investigated by using Eq. (9), which expresses the amount of melting as a function of radiation balance and air tempera- ture. The amount of freezing and the thermal regime of the ice can be calculated by the methods described by Untersteiner. [28][29] The amount of freezing is found to be a function of ice thickness and air temperature.

Budyko and Zubenok used these relationships for melting and freezing to calculate the variation in ice thickness and the equili- brium conditions for various areas, using for each area the observed annual pattern of air temperature. [53] Their results agreed well with the thickness and boundaries of observed ice cover.

By the same method was investigated the effect of summer tempera- ture anomalies on ice melting, calculations being performed for each month not only of summer but of the whole year. Budyko reports that positive summer anomalies of $4^{\circ}C$ would melt the central Arctic ice of mean thickness of 4 m in 4 years. He argues that in reality a much smaller temperature anomaly would have the same effect because the original calculations did not take into account the lowered albedo in each succeeding year, as the ice margins recede and summer thaw comes earlier. Introducing this refinement, he concluded that an

initial increase of average summer air temperature of 2°C would lead
to a substantial decrease in ice cover in peripheral seas and in a
few decades would completely destroy the pack ice.

In this connection, it may be noted that the largest observed
summer anomaly was about 2.5^{o} in 1921. According to Zubov,[35] the
pack ice disappeared from hundreds of thousands of square kilometers
in the European sector of the Arctic.

Interaction among Ocean, Atmosphere, and Pack-Ice Margin

It has long been observed that cyclones tend to follow avenues
of warmer and moister surfaces. Warm seas give off enormous amounts
of heat when exposed to cold air, thus influencing air to rise and
corresponding amounts to flow in at lower levels; that is, warm seas
tend to create mean lows above them. The North Atlantic and the
Norwegian Sea are especially effective in this respect.

In the Atlantic area between 20°N and 65°N, the mean heat exchange
between ocean and atmosphere is about 31 kcal/cm^{2} but reaches extreme
values off Nova Scotia.[54] In general the mean heat exchange is about
40 kcal/cm^{2*} at 50°N and only a little more than this farther north.

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*A value that is about equal to assumed mean surface-heat loss
from our hypothetical open Arctic (Fig. 26). Thus the magnitude of
evaporation and turbulent-heat exchange shown in Fig. 26 gives a
reasonable result.
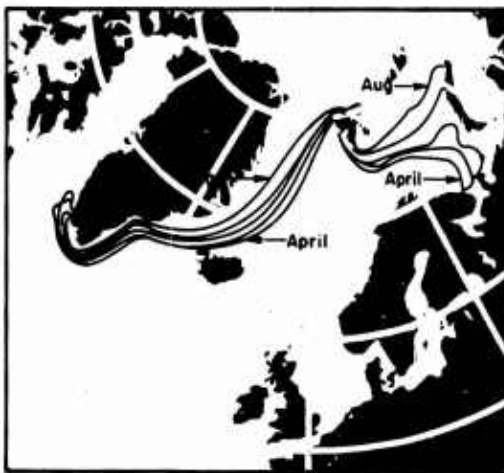
Fig. 33 -- Monthly means of pack-ice edge,
April--August (after Zubov).

Figure 33 shows the position of the pack-ice margin in the Norwegian and Barents Seas between times of maximum and minimum ice extent. During winter the open water narrows on the west but still reaches West Spitsbergen and Murmansk on the north. Thus it constitutes a long, narrow avenue of intense surface heating. Cyclones from the North Atlantic tend to follow along this ice margin.

The extent of pack ice has been observed to vary widely from year to year, and certain characteristics of atmospheric circulation seem to be associated with these changes. It has been observed that when the ice margin is far advanced, cyclones from the North Atlantic tend to take more southerly paths. When the ice margin recedes, they tend toward more northerly paths. Many attempts have been made to identify cause and effect in these relationships but without much success. Over 30 years ago (1932), Sandstrom correlated water temperatures in the North Atlantic with the displacement of the mean highs and lows in the Atlantic.[55] Later (1943) Zubov pointed out that anomalies of water temperature require several years to pass through the Norwegian Sea and into the Arctic Basin, and that the north--south temperature gradient derived from following such movements correlated well with the latitude at which cyclones pass the Kola meridian (Ref. 35, page 86).

Of course, cyclones also tend to follow the direction of upper-level wind flow, which is in turn related to the wave pattern of the zonal circulation and thus to the positions of mountain barriers and heating patterns in distant parts of the hemisphere. These many influences on storm tracks interact in such a complicated way as to defy complete analysis.

In the U.S.S.R. a great deal of effort has been expended in classifying various patterns of atmospheric circulation and correlating them with various factors affecting atmospheric heating. (This approach has not been very popular in the U.S.)

In the 1940's G. Ya Vangengeym concluded that the many observed patterns of atmospheric circulation could be classified under three fundamental types, which he called western (W), eastern (E), and meridional (C) (Fig. 34).

This scheme has since been developed in greater detail by
A. A. Girs and others at the Arctic Institute in Leningrad. For a
complete discussion of these circulation types, their frequency of
occurrence, correlation with indices of solar activity and astronomi-
cal factors, etc., the reader is referred to the recently published
memorial volume to G. Ya Vangengeym. [56]

A similar scheme of classification has been developed by
B. L. Dzerdzeevskii at the Institute of Geography in Moscow. He
identifies thirteen types of Northern Hemisphere circulation falling
into four main groups, as shown in Fig. 35. Groups I and III (in
the two left boxes in Fig. 35) are conducive to rainfall in the
United States and Western Europe, whereas groups II and IV (right-
hand boxes) are associated with aridity (as during the 30's). Group
II is a rare transitional type, whereas the others correspond
roughly to Girs's W, C, and E.

Weather patterns for the last sixty years have been analyzed to
determine the relative frequency of occurrence of each type. The
variations over this span of time have been considerable, and various
statistical correlations relate these variations to solar activity.
As yet, however, the chain of cause and effect (from solar activity
to heating pattern to atmospheric behavior) is very obscure indeed.

Figure 36 summarizes some general relations derived from this
work. In creating this figure, the mean frequency of occurrence of
each type of circulation was determined by examining each day from
1891 to 1955. These are compared with the Wolfe number, a number
proportional to the mean number of sunspots.

It can be seen that the decreasing number of sunspots from the
1890's to the 1930's was accompanied by a steadily decreasing frequency
of the meridional type of circulation. During that period, the western
type of zonal circulation occurred more frequently; the eastern type
occurred less frequently than normal for the entire 65-year period
1890--1955. Since the 1930's, the number of sunspots has increased
sharply; the meridional type of circulation has become more frequent;
the eastern type has become more frequent than normal and the western
type less frequent.
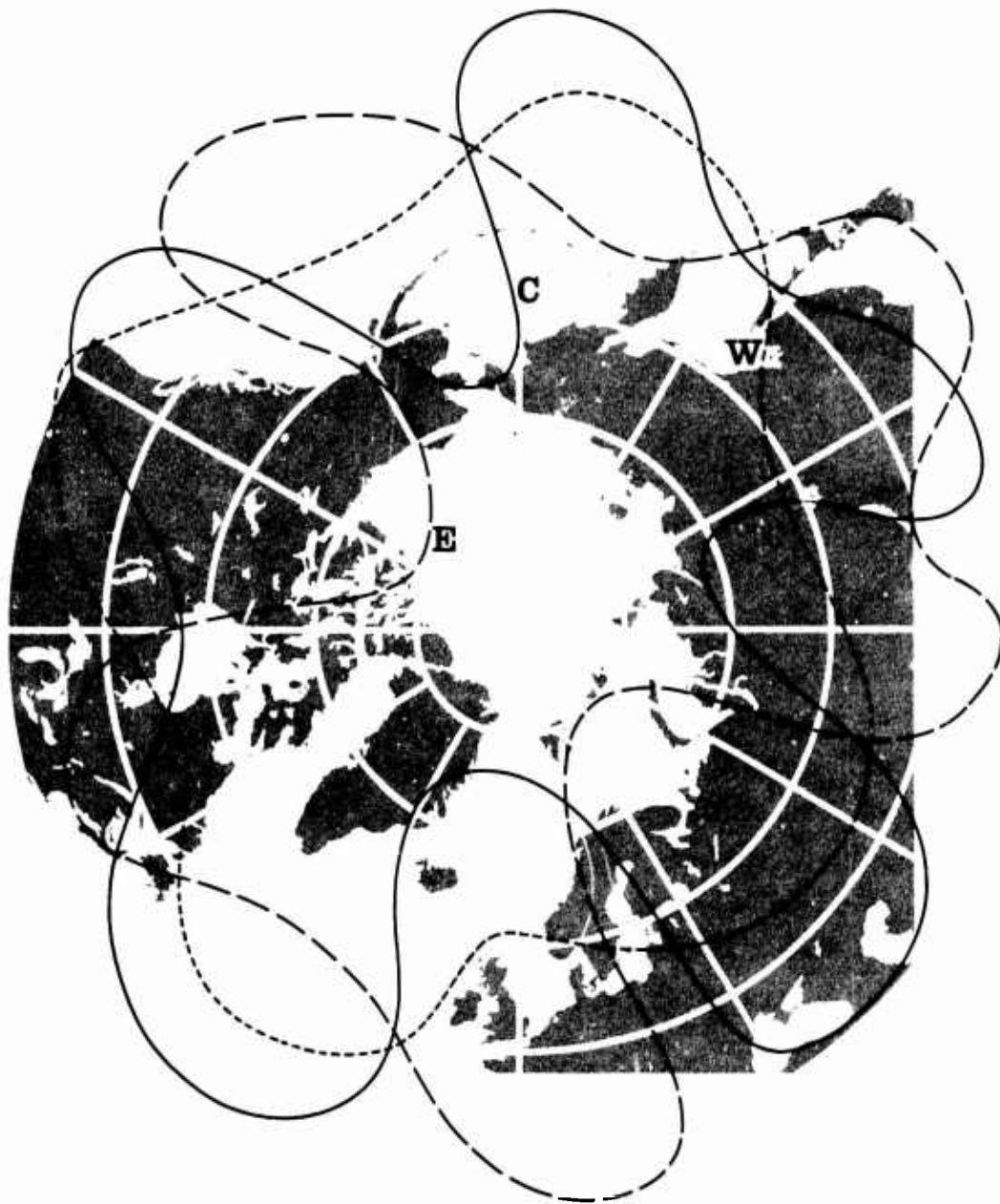
Fig. 34 -- Trough and ridge patterns of main circulation types:
E, eastern type; W, western type; and C, meridional
type (after Vangengeym and Girs[56]).

These relationships suggest that the fundamental cause of
changes in circulation types is of solar origin. The trends are most
strongly evident in the Atlantic sector of the Arctic, where as noted
previously the interaction of water temperature, ice extent, and cir-
culation can be especially sensitive to variations in solar heating.
The decrease in ice extent in the Atlantic sector between 1891 and
1940 is indicated by Fig. 37, which also shows similar evidence of
ice extent over the last thousand years. What atmospheric circulation
was like during most of these centuries is unknown.

Observed Variations in General Circulation and Ice-Extent

An immensely detailed study of weather records has been compiled
by H. H. Lamb in an attempt to identify climatic changes and their
causes. [58][59] A few of his findings that are relevant to the
present study follow.

There has been an increase in the intensity of the general
atmospheric circulation from about 1800 to about 1940 (Fig. 38). The
trend was evidently world-wide but was most pronounced in winter
over the North Atlantic. The long-term change in zonal wind speed
amounted to about 10 per cent, but shorter-term variations (one or
two decades) amounted to 20--25 per cent over the North Atlantic.

A great diminution in the extent of pack ice accompanied the
increasing vigor of atmospheric circulation, especially in the region
between Greenland, Iceland, and Norway. Lamb interprets this diminu-
tion as an effect of greater atmospheric heat advection into the Arctic
rather than as a cause of more vigorous circulation.[*]

According to Lamb:

> There was considerable variety in the forms taken by
> both strong and weak circulations in different years and
> decades. These variations may provisionally be attributed
> to variations in the thermal pattern -- as when the greatest
> recession of the Arctic ice in the 1930's opened up a depres-
> sion track in mid winter far towards the north in the

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[*]According to the usual reasoning, less ice extent should contri-
bute to a weakening of atmospheric circulation. As is pointed out
at the beginning of Section VI, the opposite conclusion is presented
here, namely, that less ice extent should strengthen atmospheric
circulation, especially in summer; thus, decreasing ice extent
enhances rather than opposes a trend toward more vigorous circulation.
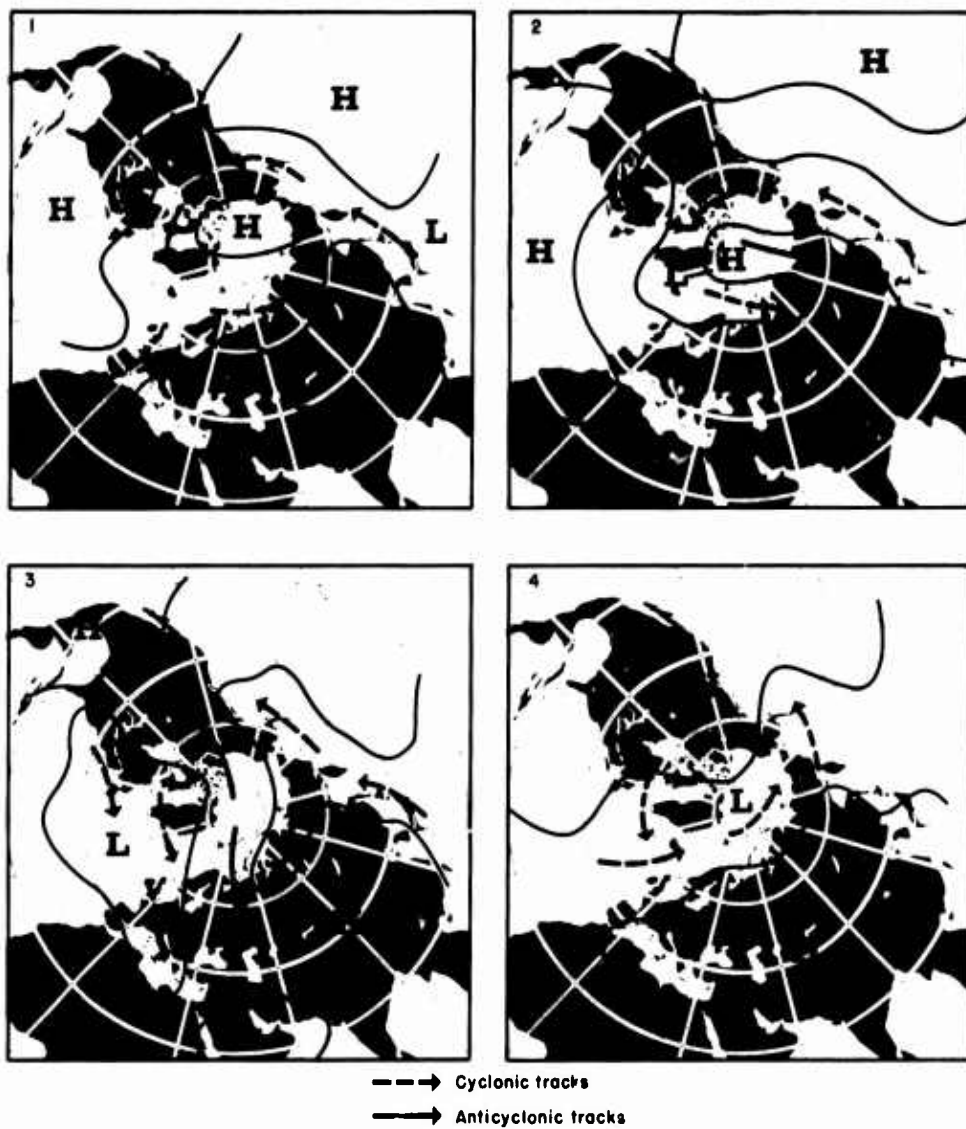
Fig. 35 -- Dzerdzeevskii's main types of atmospheric
circulation patterns.
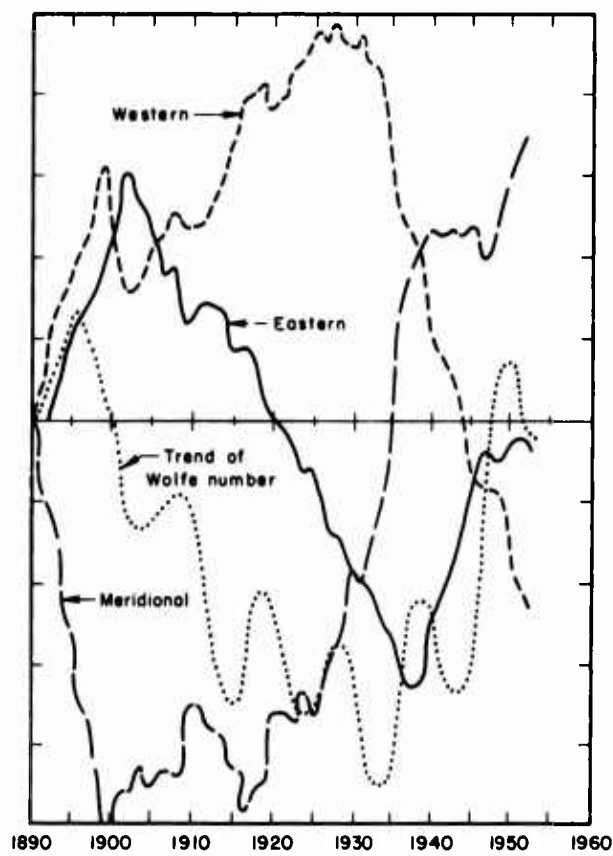
Fig. 36 -- Trends in the occurrence of basic circulation types
and in solar activity (after Eigenson, 1963).
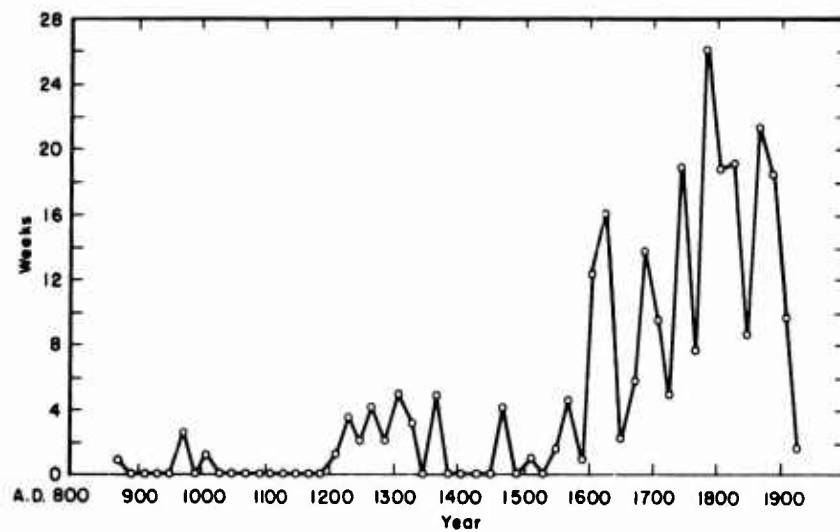
Fig. 37 -- Number of weeks with ice in the Iceland area according
to Lauge Koch (1945). The climate change in the 13th
century led to the extinction of the Greenland colonies.
Some of these sites were exposed by receding glaciers
during the sharp warming of the early 20th century.

Spitzbergen area. In this phase southerly meridional components were more prominent than zonal in the strong circulation over the northeastern Atlantic, Norwegian Sea and northwest Europe....

There are reasons to believe that changes in the latitude of the main ice limit in both northern and southern hemispheres induce some sympathetic shifts in the prevailing latitudes of the depression tracks. There are also suggestions which appear to be well founded, of parallel movement of the northern and southern ice limits since the sixteenth century. [See Fig. 39 in the present work.]

Insofar as changes of intensity and latitude of the main zones of the atmospheric circulation may be inter-linked, it appears that a weakened circulation during the cooler seasons of the year allows more ice to form on the Arctic seas and this in turn shifts the main thermal gradient towards lower latitudes, especially in the North Atlantic and neighbouring sectors.[58]

Lamb says further:

...the main depression paths, indicated by the lati-tude of lowest average pressure in January and July, were moving generally farther north during the period of strengthening of the main zonal wind systems. That was during the general climatic warming period from 1800 (or probably earlier) to about 1930. The slackening of the general zonal circulation, which seems to have set in over the last 30 years or rather more, has been accompanied by a marked tendency for Atlantic depressions to revert to more southerly tracks.... This southward shift of the depression tracks, and therewith of the zone of maximum windiness...over the ocean to the regions south of Ice-land, preceded any renewed increase of the Arctic pack-ice. (By 1938 the sea ice had retreated farther north than ever before in modern times, and those who reasoned from extrapolation alone began to speak of the possibility of an open Arctic Ocean by the end of the century.)[60]

Lamb has also drawn conclusions about variations in the dominant wavelenths in the upper westerlies and also variations in their speed. His findings regarding changes during the last two centuries are shown as Fig. 39.

He estimates that the mean speed of the upper westerlies in July has varied from about 16 m/sec during periods of maximum ice extent to 19 m/sec during periods of minimum ice extent, with a concurrent
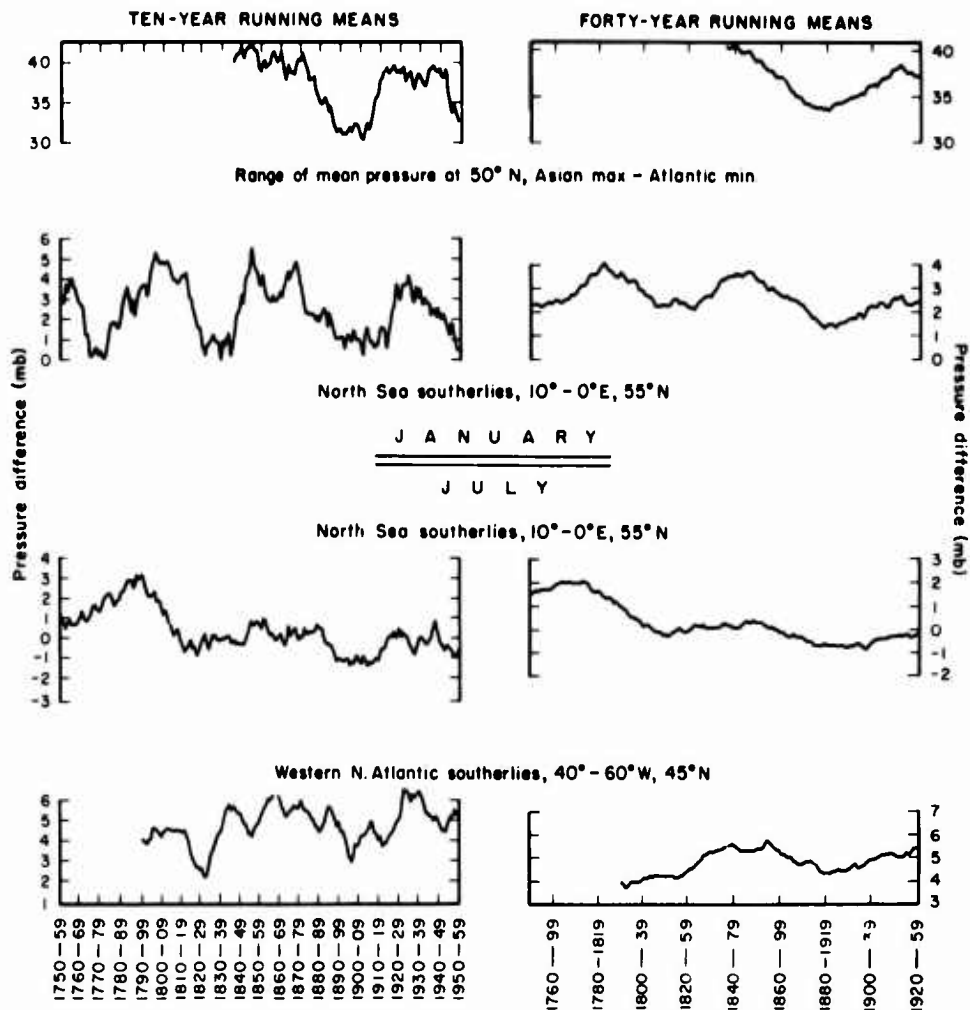
Fig. 38 -- Pressure-difference indices of circulation intensity
in January and July for two centuries (after Lamb).
These indices can be compared with those in Fig. 27.
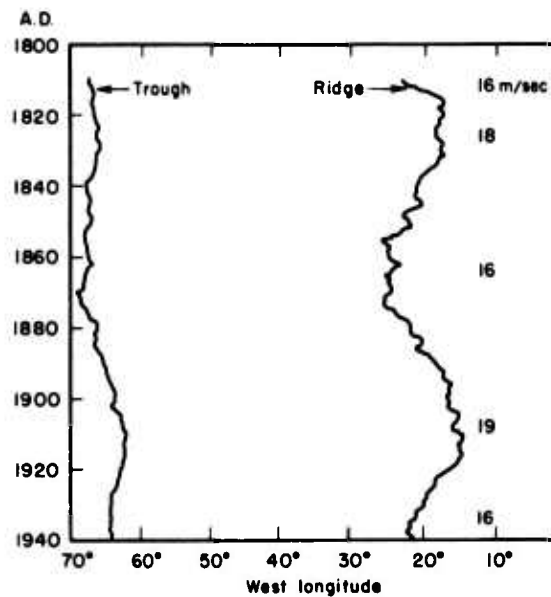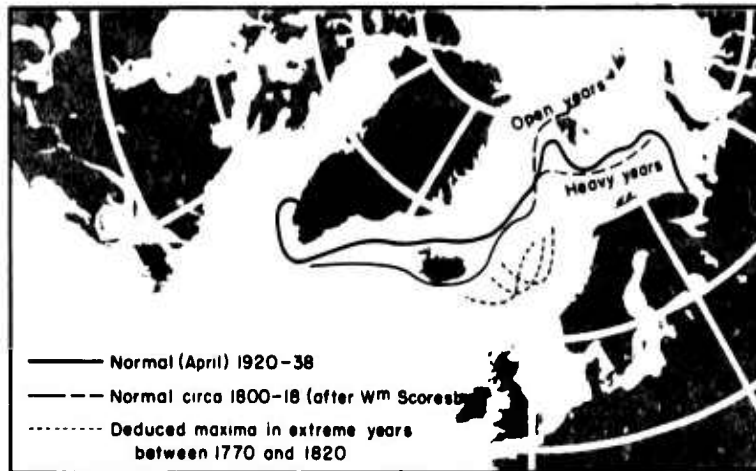
Fig. 39 -- Lamb's findings. Ice extent and variation of
speed and wavelength in the upper westerlies
in July. Above, spring maximum of pack ice
in various years. Below, longitudes of the
semi-permanent surface-pressure trough and
ridge at 55°N. Forty-year running means
are used. Speeds at the right refer to
upper westerlies.

change in wavelength in the zonal flow. It is of interest to note
that greater circulation intensity corresponding to minimum ice
extent is entirely consistent with the rates of atmospheric heat
loss shown in Fig. 27.

VII.   INTERACTION OF THE ARCTIC OCEAN WITH OTHER OCEANS

From the foregoing discussion of variations in atmospheric cir-
culation, it is apparent that the geographical pattern of surface
heating undergoes considerable year-to-year variation in the Atlantic
sector. Variations in ice extent and variations in water temperature
both seem to affect atmospheric circulation patterns, and to be in
turn affected by them. It is also in the Atlantic sector that the
principal exchange of water and heat takes place between the Arctic
Ocean and other oceans. Ewing and Donn attribute the periodic dis-
appearance of the pack ice and the subsequent ice ages to variations
in oceanic heat transport into the Arctic. There is, in fact, strong
evidence that water exchange between the Atlantic and the Arctic has
undergone wide variation at intervals comparable to glacial--intergla-
cial intervals.

To get a coherent general picture of how the Arctic heat budget
interacts with the climate of other latitudes, it is necessary for one
to include a quantitative description of oceanic heat exchange. Only
by quantitative assessment can the role of oceanic heat exchange be
placed in perspective in relation to the other factors affecting the
heat budget. This section gives the essentials of such a quantitative
description.

VERTICAL STRUCTURE AND HEAT CONTENT OF ARCTIC-OCEAN WATER

The Arctic Ocean is very different from the other oceans of the
world because (1) it is relatively isolated from other oceans, and
(2) the presence of the pack ice radically transforms its physical

characteristics. The mantle of ice protects it from the mechanical action of the wind, keeps it in relative thermal isolation from the atmosphere, and by melting and freezing processes, largely determines its vertical structure.

The vertical structure, which is relatively uniform over the central Arctic, is shown in Fig. 40. The most significant features are the extraordinary stability of the surface layer and the presence of a relatively warm layer at 300--700 m. The deeper waters are characterized by a nearly uniform salinity (almost 35 parts per thousand) and only gradual temperature changes. There are three important physical characteristics of salt water[*] that help explain these features.

One is that salt water of a salinity over 25 parts per thousand continues to become more dense when cooled to its freezing point (about $-2^{\circ}C$), unlike fresh water, which expands as its temperature approaches freezing. Thus sea water cooled from above will sink, unless prevented from doing so by salinity stratification, until the entire water body has been cooled to freezing temperature.

Second, the density of sea water is very sensitive to its salinity. One consequence is that evaporation in excess of precipitation increases the density of surface waters and induces vertical mixing. Outside the Arctic Basin, in regions of very high evaporation, such as the Norwegian Sea, this vertical mixing is so great as to create deep convection and nearly uniform water throughout the water mass. Conversely, an excess of fresh water over evaporation creates stability at the surface unless the fresh water is mixed by mechanical action of the wind.

Third, when sea water freezes, most of the salt is expelled from the solid phase and sinks as heavy brine down to lower levels. Thus when pack ice melts, it releases fresh water into the surface layers

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[*]For a detailed treatment see Zubov, Ref. 35 or "The Oceans," by Sverdrup, Johnson, Fleming.
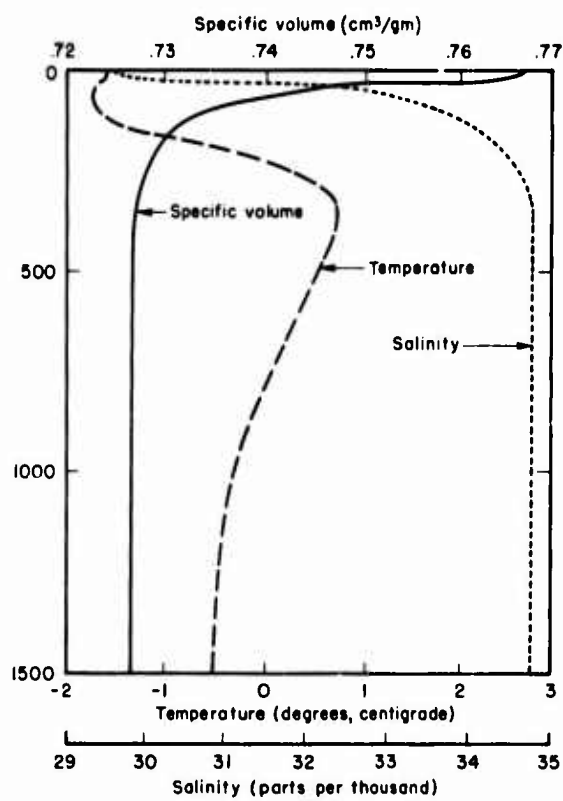
Fig. 40 -- Typical distribution of temperature,
salinity, and specific volume in
the Arctic Ocean.

The low salinity and the high stability of the surface waters
are due to a combination of fresh-water runoff from land and to the
release of fresh water by melting pack ice. Each year the export of
fresh water in the form of ice is approximately balanced by the run-
off of fresh water from land.

If the pack ice were removed, the surface layers would be exposed
to mixing by wind and, in some areas, to mixing induced by a salinity
excess at the surface due to evaporation. How deep such mixing would
reach is uncertain, since wind mixing normally is effective only in
the upper 100 m. It is nevertheless timely to note here how much heat
is stored in deeper layers of the ocean.

If we take the average temperature of the first 1500 meters as
$0°C$, the heat released by cooling this depth to $-2°C$ would be about
3×10^5 cal cm^{-2} (or 9 kcal cm^{-2} mo for almost three years). If we
also consider the amount of heat from depths below 1500 m, the above
figures would be about doubled. Thus it is apparent that a large
quantity of heat is stored in deeper levels of the ocean, but its
availability to the surface would be limited by salinity stratifica-
tion, even in the absence of pack ice.

The warm water layer between 300 and 700 m in Fig. 39 is due to
advection from the Atlantic. The manner in which this oceanic heat
is brought into the Arctic explains the ice configuration in Figs. 33
and 39 and has important climatic consequences. It is therefore
described in detail below.

WATER EXCHANGE WITH OTHER OCEANS

The water, heat, and salt exchange between the Arctic Ocean and
other waters has been studied by many investigators since Nansen's
pioneering voyage[61] of 1893--96: in the United States by Barnes
and Coachman (University of Washington)[62] and Bloom (Navy Electronics
Laboratory); in Canada by Collin (Ottawa)[63] and Vowinckel and Orvig
(McGill University); in Norway by Mosby (Bergen);[63] and in the
Soviet Union by many investigators.[63] For a brief but comprehensive
treatment in English, the reader is referred to the summaries by

Timofeyev, Panov, and Shpaiker in the April 1964 issue of Deep Sea Research. The following data are based on that source.[64][65]

The main exchange of water is with the Atlantic and is dominated by the East Greenland Current flowing south and a branch of the North Atlantic Current flowing north through the Faeroe--Shetland channel.

The East Greenland Current is the result of the general pressure and wind distribution over the Arctic and along the coast of Greenland (Fig. 28). The relatively shallow, cold, and swift current is pressed against the land barrier of east Greenland and carries out of the Arctic Ocean about 124,000 km^3 of cold water and 3000 km^3 of ice each year.

The warm North Atlantic Current (called the Gulf Stream until it passes the Grand Banks) originates as the North Equatorial Current where warm surface waters driven by easterly trade winds are deflected north to form the Florida Current and the Gulf Stream.[66] The North Atlantic Current separates the relatively warm waters of the central Atlantic from the colder waters to the north and as a boundary phenomenon is largely governed by the vertical characteristics of the two water masses it separates. It passes northeastward across the North Atlantic, and a major branch enters the Norwegian Sea near the Faeroes and Shetlands (Fig. 41).

Another current enters the Arctic via Bering Strait (30 x 10^3 km^3/yr), and 4.4 x 10^3 km^3/yr is added from river runoff. The outflow through the passages in north Canada is 40 x 10^3 km^3/yr (Fig. 41).

NET OCEANIC HEAT EXCHANGE

The transport of heat associated with this water exchange is shown in Fig. 42 and Fig. 43. For the ocean, the Norwegian Sea is an enormous heat sink. In terms of heat units* of 10^{15} kcal/yr, 1922 units enter the Norwegian Sea (Fig. 41). Seventy-three per cent of this (1410), plus an additional 415 net gain from solar radiation, is dissipated in the Norwegian Sea, 42 per cent in evaporation, 35 per cent in turbulent exchange of sensible heat with the atmosphere,

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*Referred to the temperature of water leaving the Arctic.
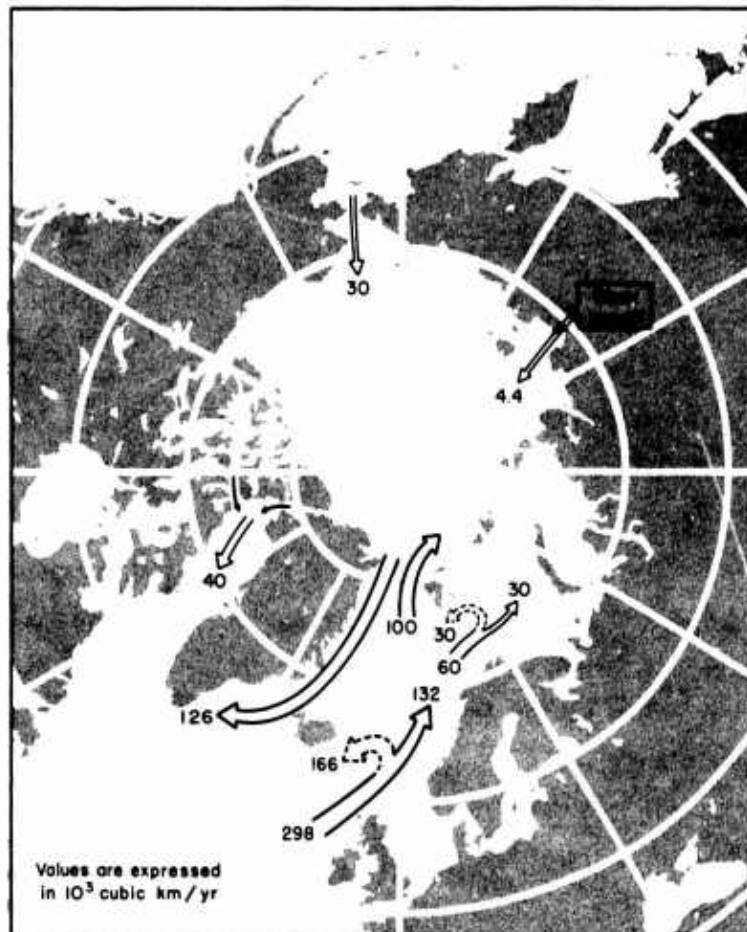
Fig. 41 -- Water balance of the Arctic Ocean.
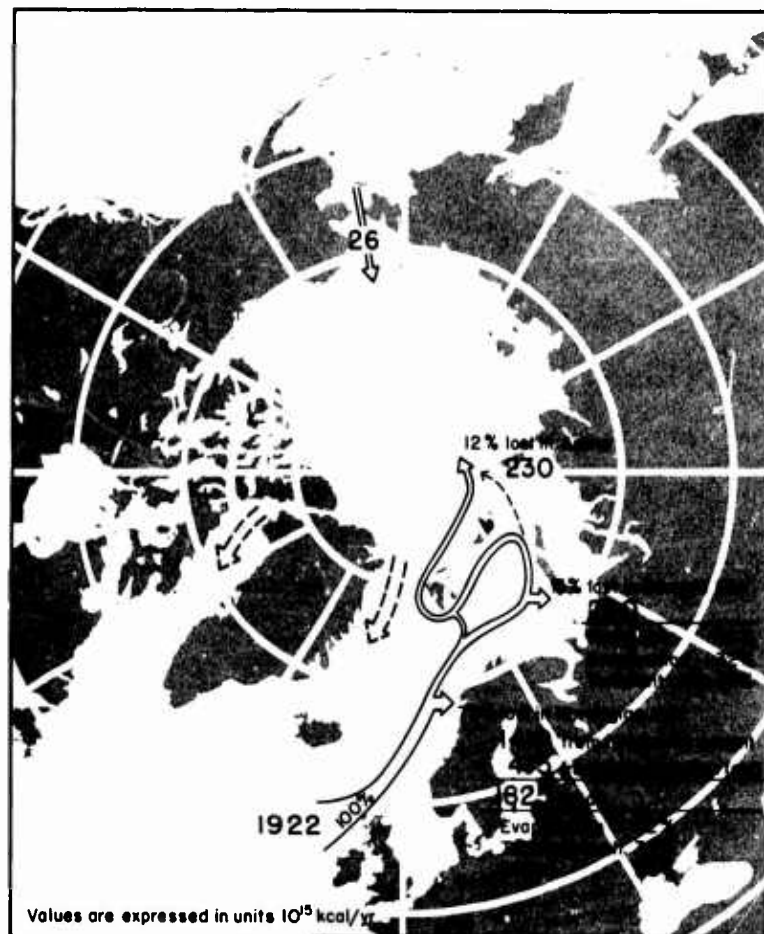
Fig. 42 -- Annual oceanic heat transport, not including
the export of ice, which also amounts to
about 250 x 10^{15} kcal/yr.

and 23 per cent by mixing with colder water. This vast heat loss to
the atmosphere is largely responsible for the winter temperature on
the Norwegian coast of about $0^{o}C$ instead of $-25^{o}C$, which is about the
mean temperature for such latitudes. Most of the evaporation occurs
in winter, adding greatly to winter rainfall in northern Europe.

Only 12 per cent of the heat entering at the Faeroes reaches the
central Arctic under present climatic conditions, while 15 per cent
enters the Barents Sea. Thus year-to-year variations in water tempera-
ture in the North Atlantic Current can greatly affect ice conditions
in the Norwegian, Barents, and Greenland Seas, but the effects in the
central Arctic are relatively small.

SEASONAL AND GEOGRAPHICAL PATTERN OF HEAT LOSS

The intensity of oceanic circulation varies greatly during the
year. Figure 43 shows the seasonal pattern of heat transport from
the Atlantic into the Norwegian Sea and from the Norwegian Sea into
the Barents Sea and the Arctic Ocean. It may be noted that water
transport is greatest in winter but that heat transport from the
Atlantic is greatest in fall, before water temperatures have cooled.

Figure 44 shows the annual variation in heat and water flow into
the central Arctic and the annual water balance. The heat flow in
September is more than five times as large as in April. The relative
importance of the Atlantic waters is apparent when we note that the
12 per cent of the heat from the Faeroe--Shetland channel that finally
reaches the central Arctic is about nine times greater than the net
heat transport through Bering Strait.

Through Bering Strait, net water flow is northward during the
year, but heat flow has a sharp maximum in summer and is negative in
winter. (The incoming water is colder than the water it displaces.)

As noted in Section IV, the mean heat flow from the ocean to the
bottom of the ice is only about 1.6 $kcal/cm^2$ yr.[*] This is a very small
component compared to the radiation components of the heat budget.
However, in the narrow area north and east of Spitsbergen that marks
the course of the Atlantic water, the annual heat loss upward may

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[*]The central Arctic comprises about 10^{17} cm^2 (it does not include
the Barents and Norwegian Seas).
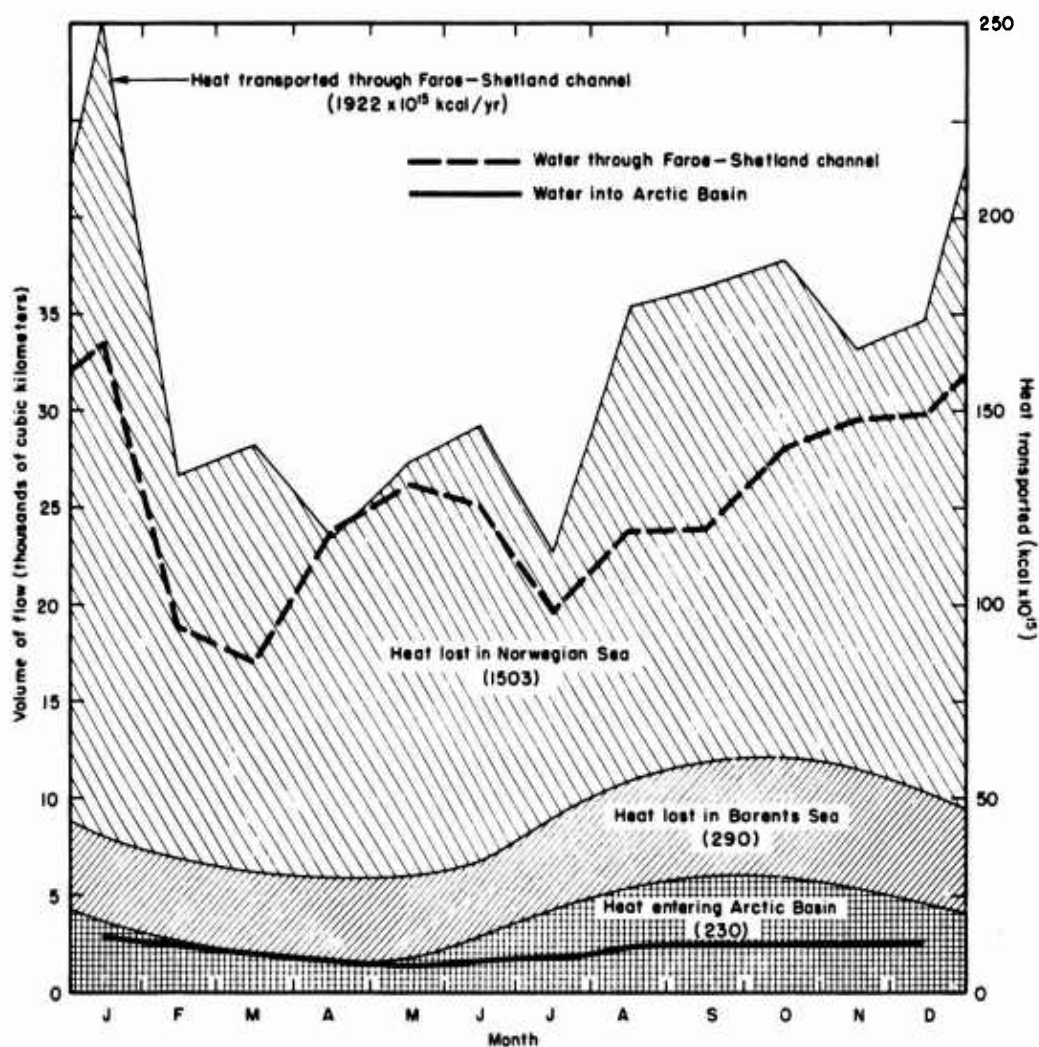
Fig. 43 -- Seasonal pattern of heat flow from the Atlantic.
Because of the high heat loss in the Norwegian
and Barents Seas, only 12 per cent of the heat
from the Faeroes reaches the Arctic Basin. The
ice cover is a very effective "lid" on vertical
heat transfer: it is self-regulating in that
more oceanic heat causes more open water and
greater heat loss upward.

reach 8--10 kcal/cm^2 (Fig. 45). This large quantity of heat is enough to affect significantly the thickness and extent of the ice in this peripheral area. Thus an anomaly in water temperature, by affecting the extent of ice cover, also brings to bear the thermal leverage represented by the differences in Fig. 21 (ice-covered Arctic Ocean) and Fig. 26 (ice-free Arctic Ocean). This change in heating pattern at the surface may bring additional thermal leverage by affecting storm tracks.

After passing west of Spitsbergen, the Atlantic water turns eastward and spreads out over the Arctic Ocean in about six years. Figure 45 shows this yearly movement. Patterns by which the heat of the North Atlantic Current is dissipated are shown in Fig. 45 and in Table 11.

Table 11

HEAT LOSS OF ATLANTIC WATER IN THE ARCTIC BASIN

(in per cent of total)

Troughs:	Nov--Jun		Jun--Oct		Total
	Per Cent up to the Surface	Per Cent down to Deeper Levels	Per Cent Up	Per Cent Down	
Nansen	38.0	4.5	19.0	2.2	63.7
Marakov	10.6	1.8	5.3	0.9	18.6
Beaufort	10.4	1.4	5.2	0.7	17.7
Total	59.0	7.7	29.5	3.8	100.0

In the narrow zone from Spitsbergen toward the east, the loss to the atmosphere reaches 9 kcal/cm^2 for the period November--June, whereas in the Pacific sector it is less than one-fourth of this amount. Thus it can be seen that whereas the total heat from the Atlantic is relatively small in the budget of the central Arctic, the Atlantic heat (including its anomalies) does strongly influence the heat budget of this sector of the Arctic.
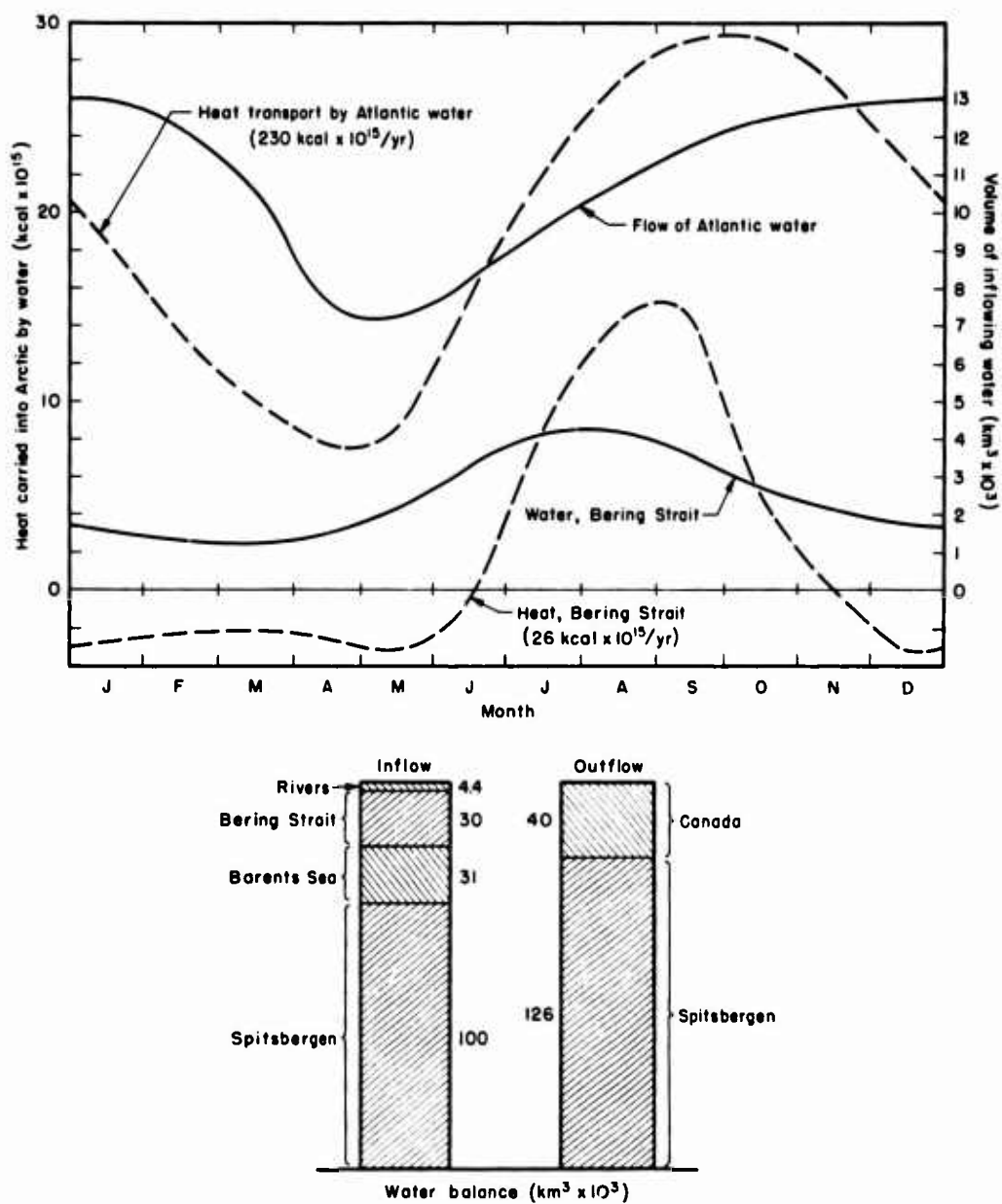
Fig. 44 -- Seasonal variation of the heat flow
into the central Arctic.
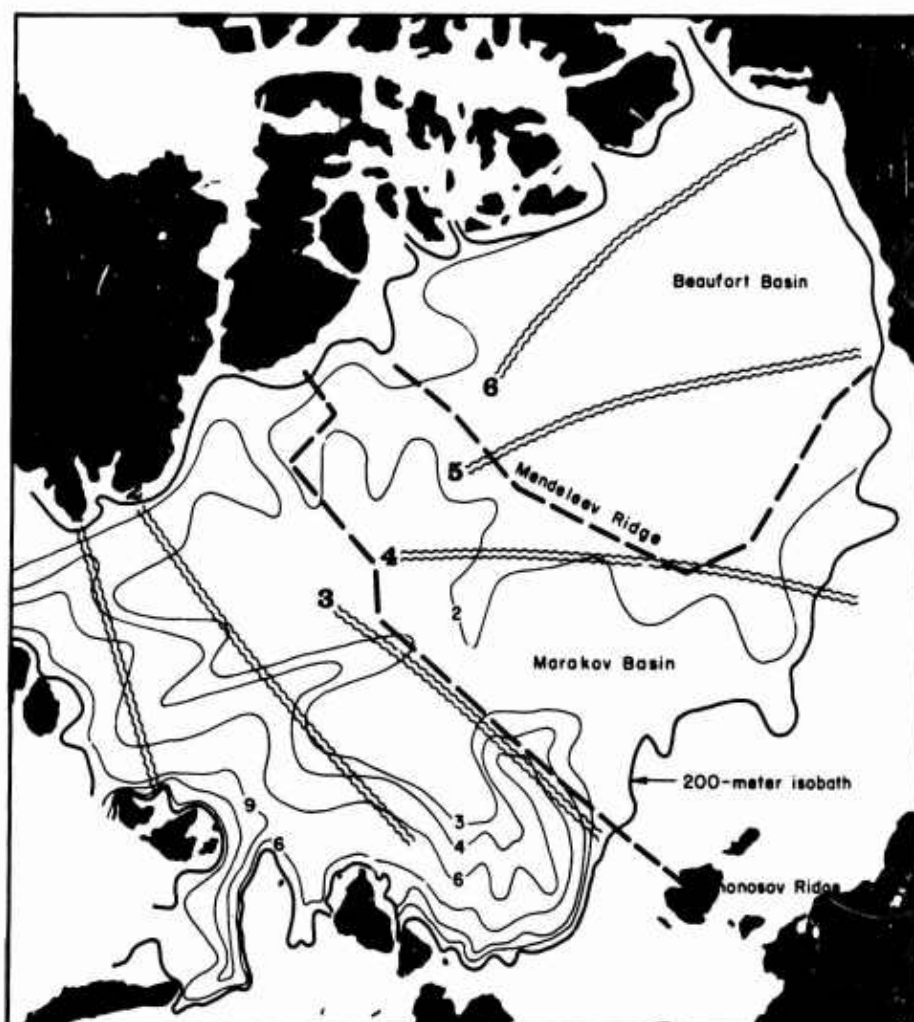
Fig. 45 -- Amount of heat given off upward by water from the
Atlantic in consequence of turbulent thermal
conductivity, November to June. Contours are
turbulent-heat emission lines in $kcal/cm^2$. Wavy
lines show year-to-year progress of Atlantic
water as it spreads over the Arctic Ocean (after
Panov and Shpaiker).

YEAR-TO-YEAR VARIATION IN OCEANIC HEAT TRANSPORT

Hydrographic sections were taken along the ice barrier near Spitsbergen during most years of the 1930's and regularly after 1950. From these data it appears that year-to-year variations in oceanic heat brought into the central Arctic amount to 15--30 per cent cf the mean value. Thus in 1939 a value of 302 x 10^{15} kcal was recorded, whereas in 1933 the value was 173 x 10^{15}, and in 1957 it was 175 x 10^{15}.

The magnitude of variations in the East Greenland Current is not known. In fact, the presence of ice makes hydrographic measurements so difficult in this region that no complete hydrographic sections are available. The volume of outflow is thus normally obtained as a residual. This is most unfortunate because the East Greenland Current is apparently the most important factor in the oceanic circulation for the Arctic.

VIII. POSSIBLE WAYS OF INFLUENCING THE ARCTIC HEAT BUDGET

In the foregoing sections, we have discussed the present heat
budget of the Arctic Basin, its relation to the climate at mid lati-
tudes, the range of year-to-year variations in heat budget and cli-
mate, and some possible climatic consequences of an ice-free Arctic
Ocean. Many of these relationships are still obscure, but several
routes to better understanding have been pointed out.

From present knowledge, a tentative conclusion is that a rela-
tionship does exist between anomalous ice conditions and climatic
anomalies at lower latitudes. Better understanding of this relation-
ship offers promise for anticipating year-to-year climatic variations,
even if the basic cause of both phenomena is external to the Arctic.

A closely related question is: To what extent might climate be
influenced by influencing the Arctic heat budget? This question has
two parts: What is the exact chain of cause and effect? Can causes
be influenced? The former question can be illuminated by a series of
climatic experiments of the type to be outlined in Section IX. The
latter question will be discussed first by identifying the thermal
levers available and considering the degree to which they might be
influenced by man.

We have seen that the presence of pack ice profoundly transforms
the thermal regime. Annual patterns of heating and cooling both the
atmosphere and the ocean are completely changed, as is the utiliza-
tion of solar energy. We will now consider how present engineering
techniques might possibly be used to influence the extent of pack ice.

THE RELATIVE IMPORTANCE OF HEAT-BUDGET COMPONENTS

The heat budget at the surface is described by Eq. (5):

$$F_r + F_s + F_\ell + F_c + F_m = 0 \quad .$$

The relative magnitude of these components decreases from left to right. The radiation balance is the largest component and is described by Eq. (1):

$$F_r = (Q + q)(1 - \alpha) - I \quad .$$

Any of these terms can be influenced to some degree, but some offer much more sensitive thermal leverage than others. The most sensitive leverage is exercised by the surface albedo, α, which can be influenced by various measures described below. Next in magnitude of effects comes artificial influence on cloud cover, which affects both the intensity of solar radiation reaching the surface, $Q + q$, and the intensity of net outgoing longwave radiation from the surface, I. Next in importance comes the effect on evaporation and turbulent heat exchange that might be induced through changes in the character of the surface. Finally comes change in F_c through interference with oceanic heat exchange (such as the much discussed Bering Strait dam). Each of these possibilities is discussed briefly in the following paragraphs.

INFLUENCING SURFACE ALBEDO

The total amount of solar radiation reaching the surface during the whole Arctic summer is about 75 kcal/cm^2 (Fig. 13) out of a possible 107 kcal/cm^2 under a clear sky. It follows that a change of albedo of only 10 per cent corresponds to a change in surface heating of about 10 kcal/cm^2. A few comparisons of this figure with other quantities show the enormous sensitivity of the heat budget to surface albedo. This is about twice the amount of heat brought into the central Arctic by ocean currents (including ice export). It is

heat enough to melt a third of the total pack ice. It is the equivalent of one million megatons of thermonuclear energy.[*]

This enormous thermal leverage is illustrated by Fig. 13, which shows that a much larger fraction of incoming radiation is absorbed at the surface during the latter part of the summer than during the early summer. This decrease in albedo is due to the collapse of the thin insulating blanket of snow and the formation of puddles on the ice.

Through much of the Arctic Basin the collapse of the snow and formation of puddles does not occur until after the solar-intensity maximum at the surface; in Fig. 13, the maximum of absorbed heat occurs in July, whereas the solar intensity is greatest in May and June.

It is plain that intervention to advance the time of the onset of puddling could have enormous effect. How might this be done?

The most common proposal is to distribute some dark material, like soot, over the surface. Actually such a scheme involves great difficulties for two reasons: First, the amount of material needed would require an immense logistic effort,[67] and second, such materials quickly lose their effectiveness because they melt into the ice in small vertical columns and soon form a dark layer several inches below the surface, with a relatively high-albedo surface above. The latter effect also applies to dirt or ash that might result from thermonuclear detonations or volcanic action.

The possibility of using microorganisms should not be overlooked. "Blooms" of reddish microflora have often been observed over substantial areas of pack ice near shore. Such an agent might have the advantages of staying at the surface throughout the season, of spreading itself over larger areas, and of surviving from season to season. It is possible that dispersion of the spores of microflora by aircraft would be more effective than spreading inorganic materials. Investigation of such possibilities has been alluded to by Budyko, but no specific work is known to this author.

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[*]One megaton is, by definition, 10^{15} cal.

Another scheme sometimes proposed is to reduce cloudiness by widespread cloud seeding. Actually, as is shown in Fig. 15, less cloudiness would not increase the surface radiation balance until after the albedo had decreased.

A more feasible scheme might be to attack the snow cover directly. For example, suppose that on the 1st of April, 25 evenly spaced 1 MT devices were detonated at an ocean depth calculated to create maximum wave disturbance.* Several of the resulting effects would tend to hasten the collapse of the snow cover and the onset of puddling: The momentary inundation would change both the albedo and the thermal properties; the increased mobility of the ice due to fracturing would enhance the work of the wind and increase the amount of open water.

Though it is not feasible to calculate the net result of these interacting and cumulative effects, a look at Fig. 13 shows the immense thermal leverage that is possible. If the summer-albedo change were increased (say from .8 to .4) and were advanced by only one month (say from one week after the solstice to three weeks before the solstice), the additional heat absorbed during that month alone would amount to about 8 kcal/cm^2 -- 32,000 times as much energy as the 25 megatons expended. By such means the amount of ice could be substantially reduced.

It should be noted that the very same operation carried out during winter would have the opposite effect on the ice budget. Reducing the thermal barrier of pack ice would greatly increase heat loss from the surface in winter and thus increase the production of ice. For example, only 1 per cent of the total area in open water from October to May would about double the winter's production of ice.

We may therefore conclude that while present knowledge does not allow confident assessment of climatic effects, there is reason to believe that present engineering capabilities could influence the Arctic's heat budget in either direction in amounts that have climatic

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*As a rough estimate, for one MT in an open ocean, a wave height of 60 ft at 100 mi is reasonable.

significance. For example, Soviet scientists estimate that changing the albedo of the surface by only 10 per cent would melt the pack ice in 8--10 years, and a change of 20 per cent would remove the pack ice in 2--3 years.[68]

INFLUENCING CLOUD COVER

In recent years we have learned that under certain natural conditions we can dissipate clouds or influence their development by the use of insignificant amounts of appropriate reagents such as silver iodide. Clouds can sometimes be formed in saturated clear air by the introduction of condensation nuclei; on the other hand super-cooled water can be dissipated by introducing nuclei.

It is characteristic of the Arctic that certain features of the thermal regime persist over vast areas. For example, the low, thin stratus of summer persists to 80--90 per cent in time and space, is very uniform in physical character, and is almost ideally subject to dissipation by seeding. Techniques for stratus dissipation have been widely published and will not be discussed here. It is of more immediate relevance to note how changes in cloud conditions can change the heat budget.

In general, clouds in the Arctic increase the longwave-radiation balance of the surface. If the cloud tops are colder than the surface (as for high clouds in summer) the total radiative loss to space is reduced. If the cloud tops are warmer than the surface (medium or low clouds in winter) the loss to space is increased.

The magnitudes of these effects are substantial. For instance, cloud tops introduced at 500 m in July would decrease outgoing long-wave radiation to space by only 0.15 $kcal/cm^2$ mo over clear sky conditions, but cloud tops at 2500 m would reduce it by 1.2 $kcal/cm^2$ mo and at 5000 m by 3 $kcal/cm^2$ mo.[11]

The present cloud regime reduces loss by only about a third of the latter amount. Thus more high cloudiness in summer could reduce longwave loss to space.

At the same time summer cloudiness reduces the solar flux passing through the atmosphere to the surface. Low stratus in summer reduces solar flux by 20--30 per cent and has no appreciable effect on the quantity of outgoing longwave radiation. Comparing these influences, it appears that thin, high cirrus above the present low stratus would increase the summer heat income, but dissipation of the summer stratus would not improve it until the albedo had been lowered.

In winter the cold surface under the deep inversion makes radiation loss much less than in summer. Thus a low cloud cover introduced in winter would actually increase loss to space, and medium clouds would have little effect.

From the foregoing it can be seen that changes in cloud cover are significant but less effective in altering the heat budget than are changes in surface albedo. The most effective way to use clouds as thermal levers would probably be to dissipate low stratus in July, when surface albedo has been lowered and solar flux is still high, or to create high clouds in September, when solar flux is small but longwave loss is still high because of the warmth of the surface.

INFLUENCING EVAPORATION AND SENSIBLE-HEAT FLUX

Evaporation and sensible-heat flux are normally calculated by simple empirical formulas of the form previously discussed. Laikhtman, however, has developed more sophisticated formulas, which relate more closely to the physics of the boundary-layer processes.[32] Using these formulas, Laikhtman has calculated the changes in evaporation and turbulent heat exchange that would result from a two-degree change in air temperature or a 10 per cent change in humidity, wind speed, or radiation balance at the surface. His results are shown in Table 12.

From these data, the variations both in air-mass temperatures and in the radiation balance of the surface are clearly important to the heat budget of the surface. During changes in humidity of the air mass there is a mutual compensation of the changes in turbulent heat exchange and heat of evaporation, and these tend to cancel each other.

Table 12

CHANGES IN EVAPORATION AND TURBULENT-HEAT EXCHANGE AS A FUNCTION
OF AIR-MASS TEMPERATURE, RADIATION BALANCE, AND WIND SPEED

Month	Air-mass Temp. $dT_a = -2^{\circ}C$		Rad. Balance $dR = 0.1R$		Humidity $dF = -0.1F$		Wind Speed $dv = -0.1v$	
	dP	dLE	dP	dLE	dP	dLE	dP	dLE
Jan	0.21	0.02	-0.63	-0.1	0.31	0.32	0.16	-0.13
Apr	0.48	-0.08	-0.28	-0.11	-0.52	0.54	0.23	-0.25
May	0.88	-0.71	0.97	0.51	-0.75	0.77	0.20	-0.21
Sep	0.33	-0.09	0.0	-0.03	-0.96	1.0	0.16	-0.16
Nov	0.13	0.05	-0.54	-0.05	-0.50	0.52	-0.08	0.10

While these components of the heat budget might be influenced
through radiation components, no direct way of appreciably influencing
them in the central Arctic is apparent.

WAYS OF INFLUENCING THE ADVECTION OF HEAT BY THE OCEAN

Many schemes for influencing or explaining climate are based on
oceanic heat advection. C. E. P. Brooks and N. N. Zubov considered
the effect of reduced access of Atlantic water to the Arctic, and in
recent years the ice age theory of Ewing and Donn has attracted
attention. Also the idea of a Bering Strait dam has attracted much
comment in the press.

In a way, this emphasis is ironical, for as noted previously,
oceanic heat advection is by far the smallest of the heat-budget
components in the central Arctic, and one of the more difficult to
influence. Its total annual value is comparable to the heat of only
a 3 per cent change in summer albedo.

There are several reasons, however, why oceanic heat advection
is more important to the heat budget than its net value would suggest.
One is that it is mostly concentrated in a narrow region (Fig. 45)
and hence can influence the extent of pack ice in this region and gain

thermal leverage by changing the radiation balance. Another reason
is that heat from this source can be stored in the deeper layers of
the ocean year after year. In fact, Zubov reports that the warmest
layer of Atlantic water (about 400 m in Fig. 40) increased in tempera-
ture by about 0.5°C between 1895 (Fram soundings) and 1938 (Sedov
soundings). A third reason is that great irregularities in the in-
flux of warm Atlantic water do normally occur.

One scheme that has been studied by Soviet scientists would
increase the heat flow to the Arctic by reducing cooling of the sur-
face in the Norwegian Sea. This would be done by spreading a film
of cetyl alcohol on the surface to suppress evaporation.[69] According
to Soviet estimates, the waters entering the Arctic from the Norwegian
Sea could be warmed by a combination of cloud dissipation in summer
and suppression of evaporation in winter by monomolecular films. The
net effect would be that water entering the Arctic would be raised in
temperature by 3 or 4°C, and the pack ice would be eliminated in a
few years.[42][68]

The Bering Strait dam proposal[70] is a scheme for doubling the
inflow of warm Atlantic water by blocking the inflow of Pacific water
through Bering Strait and also pumping 500 km^3/day from the Arctic to
the Bering Sea. There are many objections to both the operational
feasibility and the basic theory behind this scheme.

These objections have been most eloquently expressed by Droga'tsev,
Rakipova, and other Soviet scientists;[68][71] they will not be
repeated here. It does seem appropriate to add, however, that there
seem to be easier ways to influence the Arctic heat budget.

CONCLUSION

In view of these many possibilities and the rough quantitative
estimates of their effects, we may conclude that feasible ways can
probably be developed for influencing the Arctic heat budget in either
direction. At present it does not seem possible to anticipate the
climatic effects that might be brought about by such actions. In the
next chapter we will consider ways of improving our understanding of
"causes and effects."

IX. THE FORMULATION OF CLIMATIC EXPERIMENTS

HISTORICAL PERSPECTIVE

The basic approach underlying one of the most exciting modern developments in atmospheric sciences was succinctly stated by V. Bjerknes in 1904 as follows:[72]

> If it is true, as every scientist believes, that subsequent atmospheric states develop from the preceding ones according to physical law, then it is apparent that the necessary and sufficient conditions for the rational solution of forecasting problems are the following:
>
> 1. A sufficiently accurate knowledge of the state of the atmosphere at the initial time.
>
> 2. A sufficiently accurate knowledge of the laws according to which one state of the atmosphere develops from another.

> I
>
> The determination of the state of the atmosphere at the initial time is the task of observational meteorology....

> II
>
> The second problem then arises as to whether we know, with sufficient accuracy, the laws according to which one state of the atmosphere develops out of another.
>
> The atmospheric processes are of a mixed mechanical and physical nature. Each one of these processes can be expressed in one or more mathematical equations according to mechanical or physical principles. We have sufficient knowledge of the laws according to which the atmosphere develops if we can set up as many independent equations as there are unknown quantities. From a meteorological point of view, the state of the atmosphere is specified, at an arbitrary time, if we can determine for that time at each point, the velocity, density, pressure, temperature and humidity of the air. The velocity, as a vector, is given by three scalar quantities, the three velocity components, and one must therefore deal with seven unknown quantities.

To compute these quantities we can set up the following equations:

1. The three hydrodynamical equations of motion. These are differential relations between the three velocity components, the density and pressure.

2. The continuity equation, which expresses the principle of the conservation of mass during motion. This equation is again a differential relation between the velocity components and the density.

3. The equation of state of atmospheric air, which is a relation in finite form between the density, pressure, temperature and humidity of a given mass of air.

4. The two laws of thermodynamics, which allow us to set up two differential relations giving the rates of change of energy and entropy during the changes of state which are taking place. These equations introduce no new unknowns into the problem, as the energy and entropy are expressed by the same variables which appear in the equation of state and connect the changes of these quantities with other quantities considered as known. These other quantities are, first, the work done by the mass of air, which is determined by the same variables that appear in the dynamical equations; and secondly, the amount of heat given up or received by the mass of air, which is determined by the physical data on incoming and outgoing radiation and on conduction where the air is in contact with the ground....

III

Of the seven equations, only one, the equation of state, has a finite form. The other six are partial differential equations. Of the seven unknowns, one can be eliminated with the aid of the equation of state, and the problem then becomes the integration of a system of six partial differential equations with six unknowns, and with the utilization of initial conditions as given by the observations of the initial state of the atmosphere....

IV

The link which ties the hydrodynamic and the thermodynamic problems together is very easy to cut, so easy indeed that the theoretical hydrodynamicists have fully used it to avoid every serious contact with meteorology; for the connecting link is the equation of state. If we suppose that temperature and moisture do not enter into this equation, then we come to a "supplementary"

equation, used ordinarily by the hydrodynamicists, which is a relation only between density and pressure. Thereby one is led to the study of fluid motions under such circumstances that each explicit consideration of the thermodynamic processes automatically falls away.

Instead of making the temperature and the moisture disappear entirely from the equation of state, we can regard them, for short time intervals, as given quantities, with values derived from the observations or from the preceding calculations. When the dynamical problem for that time interval is solved, then one computes afterwards new values of temperature and moisture according to purely thermodynamical methods. These one regards as given quantities when one solves the hydrodynamic problem for the next time interval, and so on....

For about a half century after V. Bjerknes's formulation of the problem, the complexity of the full system of equations defied simultaneous solution. Climatology remained a descriptive science, and dynamic meteorology followed a pattern of analyzing one or more of the basic equations in relation to more restricted physical formulations -- supplemented empirically by observed relationships among meteorological variables and by various simplifications in the mathematical expressions.

In 1922 L. F. Richardson made more detailed proposals for numerical integration of the equations of atmospheric motion, but the practical difficulties proved to be too great to permit application of this method.

During the 1930's and early 1940's methods for forecasting future states by iterative solution of the "primitive equations" (by graphical means) began to take shape, both in the United States (Starr, Neiburger) and in the Soviet Union (Kochin, Dorodnitsyn, Blinova, Kibel). It was during and after World War II, however, that the great expansion and improvement of atmospheric observations began to fulfill Bjerknes's stated task for observational meteorology and that developments in computer science created opportunities for handling the full system of equations. Under the initial leadership of the late John von Neumann, the advances have accelerated since 1950, leading, during the mid 1950's, to the introduction of machine

forecasts for operational purposes and, in the early 1960's to numeri-
cal models of atmospheric circulation that can be used for "climatic
experiments." That is to say, the experimenter can impose an arbitrary
pattern of heating and cooling and can then let the "model atmosphere"
find its own adjustment to the given boundary conditions.

Already there are at least three operating models in the United
States that are suitable for climatic experiments, and there will
soon be more.*

In these models the "primitive equations" are replaced by their
finite difference analogues and the difference equations are inte-
grated numerically with the aid of high-speed digital computers.

At this stage the models are still rather crude, but they do
simulate atmospheric behavior with enough fidelity to show much promise
of further advances.[73] It should be noted that,for exploring cause-
and-effect relationships, such models offer important advantages over
the real atmosphere even if we were able to observe the real atmos-
phere in detail and to impose the given "cause." One advantage is the
power of the experimenter to hold all the other boundary conditions
constant while changing only the one whose effect is being investigated.

At present there remain a number of difficulties, both mathemati-
cal and physical. The experimenter is faced with a difficult trade-
off of machine time, grid-point interval, number of variables to be
carried, and time interval between iterations. However, when we con-
sider the future advances in computers already assured and the advances
in problem formulation normally expected, there can be no doubt that
opportunities for "experimental climatology" are at the threshold.
It is timely to consider which experiments would be most revealing.

ARCTIC RESEARCH AND CLIMATIC CHANGE

The climate of the earth has remained relatively uniform during
the last 1000 million years, except for brief "ice ages," during which
extensive continental ice sheets formed, advanced, and dissipated

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*Those of C. Leith (Livermore Radiation Laboratory); Y. Mintz,
(University of California, Los Angeles); and J. Smagorinsky (U.S.
Weather Bureau); and forthcoming, the model at the National Center for
Atmospheric Research.

several times. Except during ice ages, the global pattern of climate
was more equable than at present, warmer in polar regions and cooler
in the tropics. One such ice age occurred about 600 million years
ago. A series of ice ages began about 1 million years ago that may
not yet be ended. It was only some 15,000 years ago that the last
ice sheet began its retreat from the northern half of the North Ameri-
can continent and only 10,000 years ago it still reached as far south
as the Great Lakes. These glacial variations were accompanied by
great variations in rainfall and temperature patterns over the conti-
nents at mid latitudes.

Our knowledge of the pattern of these climatic variations and
their possible causes has been recently reviewed by Schwarzbach, who
sums up as follows (Ref. 74, p. 261):

> This much is certain: major climatic variations are
> extraordinarily complex phenomena, which are achieved through
> the workings of countless individual factors. It is, there-
> fore, no wonder that not one of the many hypotheses which
> have been advanced offers a complete solution; in fact, it
> still appears almost impossible to solve the mystery of the
> Ice Ages.

We have strong reasons for solving this mystery, for the pattern
of our exploding world population and its socio-economic framework
is closely related to climate. Variations of climate of the sort
witnessed even during recent decades will have increasingly grave
consequences as the supply of natural resources becomes more strained.

At the same time, the scale of man's interference with the natural
environment is approaching the level where inadvertent climatic effects
of worldwide impact could be triggered. These facts, taken together,
are enough to lend urgency to the task of unraveling these "extra-
ordinarily complex phenomena, which are achieved through the workings
of countless individual factors."

To unravel the complicated chains of cause and effect we need
working models on which simple propositions can be tested. There are
obvious ways to get on with this task.

The foregoing sections describe a "thermal model" of the Arctic
Basin that is consistent, related to climatic factors, and subject to

testing. There are two important reasons for having such a model,
even if it has large uncertainties.

One is that the climatic consequences of varying certain
parameters can be tested, thus yielding clues to cause and effect and
calling our attention to new relationships to look for in nature.

The other is that the relative importance of the factors making
up the model and our uncertainties about them are brought into focus,
thus leading to closing observational gaps and directing field efforts
along more productive lines.

CLIMATIC EXPERIMENTS WITH THE ARCTIC HEAT BUDGET

Only since the International Geophysical Year and the publica-
tion of such works as Budyko's "Atlas of the Heat Balance of the
Earth"[12] have suitable quantitative data become available to des-
cribe the heat processes over the earth's surface in a form easily
used by the experimenter. Unfortunately, Budyko's atlas, like other
sources, does not cover the Arctic Basin. In spite of this lack of
quantitative data, however, the experimenter must insert an estimate
of the heating pattern. Moreover, the experimenters are inevitably
drawn to anomalies in the Arctic heat budget. As the models are
improved they will be drawn to anomalies that are yet more sophisti-
cated.

Sections II through VI describe a thermal model of the Arctic
that is good enough to permit formulating some meaningful experiments,
even with present crude numerical models of atmospheric circulation.
Among the more obvious experiments are the effects on atmospheric
circulation of:

1.   An ice-covered Arctic Ocean without continental ice sheets
(the present climate). (See Fig. 13 and Fig. 21 in the present
work.) This would be the "control condition."

2.   An ice-free Arctic Ocean. (See Fig. 26; also Figs. 16, 17.)

3.   An ice-free Arctic Ocean with extensive continental ice
sheets.

4.   An ice-covered Arctic Ocean with continental ice sheets.

5.   An abnormal extent of ice cover. (See Fig. 39.)

6. An abnormal date of summer thaw over the whole pack.
(See Fig. 13.)

Even with improved computers and improved models, the investment
of machine time in such a program will be large and costly. It is
therefore doubly important that the "control condition" (the present
climate) be modeled as accurately as possible, for all the succeeding
experiments will yield "differences" from the "control condition."
Thus the first task is to formulate the best possible thermal model
of the Arctic and to bring about any short-term improvements in the
model that may be possible, while also undertaking a longer-term pro-
gram aimed at reducing major uncertainties.

The second task is to formulate the quantitative inputs to the
experiments outlined above in the most realistic way, considering the
limitations of the models. As a start, the values presented in the
foregoing sections would suffice.

A third task is to undertake a long-term program of field observa-
tions whose objectives are to describe the heat budget and the
detailed climatic history of recent glacial ages.

Unraveling the cause-and-effect relationships of climatic change
offers promise of learning how to anticipate climatic variations.
But that is not all. From Section VIII it is clear that even present
engineering is adequate to influence the heat budget significantly and
thereby affect climatic processes. If these processes can be under-
stood, such influences can be purposeful. And they must be purpose-
ful, for the six billion people of 35 years hence (this includes most
of those reading these lines) may well need both to maximize and to
stabilize their climatic resources.

X. NEEDED RESEARCH ON THE ARCTIC HEAT BUDGET

In Section VI, we noted that a careful quantitative assessment
of the Arctic heat budget is necessary to permit drawing conclusions
about climatic effects. For example, according to the assessment
presented here, we arrive at the unexpected and significant conclusion
that less pack ice contributes to more intense atmospheric circula-
tion rather than less intense, as would seem logical from qualitative
reasoning.

It must be emphasized, however, that there still exist some
large uncertainties in our knowledge of the Arctic heat budget --
uncertainties about both mean conditions and year-to-year variations.
For example, the most uncertain factor seems to be the amount of
solar radiation absorbed by Arctic clouds, especially summer stratus.
Should this amount turn out to be only half the values used in our
assessment -- and this is possible -- the significant climatic con-
clusion mentioned above would be questionable.

A principal aim of this study is to put some of these factors
into perspective -- to call attention to specific needs for data
and to specific opportunities for obtaining data -- to say to those
who would experiment with atmospheric circulation, "This is what we
now know about the Arctic heat budget," and to the Arctic scientist,
"This is what we need to know."

The latter conclusions follow directly from the preceding dis-
cussion but are summarized in the following paragraphs.

In some areas, it appears that rather straightforward investi-
gations could bring about major improvements in our understanding of
the Arctic heat budget. The first four programs suggested are in the
category, "critically needed and straightforward." The fifth and

sixth programs are critically needed but not easy to carry out.
Finally, are listed a number of investigations that are relevant and
needed but do not seem to warrant being assigned the same degree of
urgency.

1. Absorption of Solar Radiation by the Atmosphere

The author believes the absorption of solar radiation by the
atmosphere to be the largest uncertainty in the Arctic heat budget.
As noted in Sections II and III, Dolgin, et al., give a value of about
37 per cent for the central Arctic, while Budyko gives about 17 per
cent for the earth as a whole, and Houghton gives about 19 per cent
for the Northern Hemisphere. Vowinckel follows Houghton.

The high albedo of the pack ice and the scattering properties of
the widespread summer stratus combine to create a condition in which
absorption in the lowest kilometer of the atmosphere could theoretically
be relatively great. However, scattering and absorption by cloud
droplets is insufficiently understood to permit drawing a reliable
quantitative picture by theoretical means. As a minimum, detailed
measurements of solar-radiation intensity upward and downward are
required above, in, and below, the stratus. Also, the spectral distribu-
tion of these fluxes should be determined.

For such measurements, both balloons (or kites) and aircraft have
peculiar merits and should probably be used in concert. Balloon instru-
mentation of the type used by Kondrat'yev[20] can be used from the sur-
face during low-visibility conditions that would preclude the use of
aircraft. Since the upper surface of the stratus is usually below
0.5 km, it may be possible to use kytoons or kites. On the other hand,
it is important to obtain measurements of mean conditions over an area
and to observe the condition of the upper cloud surface and the extent
of other clouds. For these observations, aircraft are required. A
well planned program combining both techniques is needed. Instrumental
and operational problems appear to be quite straightforward.

2. Variations of Mean Albedo of the Pack Ice During Summer

As noted in Sections II and VI, an abnormality in mean summer
albedo of only 10 per cent corresponds to about 7--8 kcal/cm^2. As can
be seen from Fig. 27, this is about 10 per cent of the annual heat

advection by the atmosphere, and since this heat advection takes place mostly during the winter half-year, the consequences should be reflected in the winter circulation pattern. Moreover, year-to-year variations in annual radiation balance could be of the order of 10 kcal/cm^2, according to results from the various drifting stations. Detailed albedo observations from the Soviet drifting stations have been summarized by Chernigovsky. Large year-to-year variations are apparent, depending on the duration of the snowless period. For example in 1954 North Pole-3 had only 6 days of snowless period but in 1955 North Pole-4 had 25 days. The Soviet measurements have been quite detailed but have been largely confined to the immediate areas of the stations.

Direct measurements of mean surface albedo over the Arctic Basin as a function of both area and season are badly needed, for surface albedo is one of the most influential and most variable factors in the heat budget. In the long run, such measurements are likely to be a continuing requirement, for we will want to keep the Arctic heat budget under observation and, if possible, anticipate anomalous weather patterns.

It is ironical that systematic measurements of albedo have never been performed by United States personnel in the Arctic, even during the International Geophysical Year, although routine weather reconnaissance has been flown over the pack ice for almost 20 years. This is a relatively simple measurement and covers a vital component of the heat budget. Both the Air Weather Service reconnaissance program and the Navy "Birds Eye" ice reconnaissance offer opportunities for collecting such data.

It may be noted that satellites in polar orbits may be able to monitor both the albedo of the surface and atmospheric absorption, provided an adequate basis is established for interpretation of the data. A comprehensive summer measurement program utilizing aircraft and balloon measurements might provide such a basis. (Nimbus II is scheduled for fall 1965; if it continues to operate, a coordinated program could be planned for summer 1966.)

3.   Computation of Atmospheric Heat Advection into the Arctic

Study of Fig. 27 reveals not only the climate-forming influences of the Arctic heat budget but also the value of direct computation of heat advection as a check on the heat budget of the region. It also illustrates how remarkably few data of this kind are available, in spite of the fact that, since World War II, a reasonably good radiosonde network has been in operation around the Arctic Basin. One reason, of course, is that the reduction of such data to the simple values shown in Fig. 27 is a very laborious task. Limited attempts have been made by Vowinckel,[31] but no comprehensive computation has yet been performed. As a minimum, computations should be made for all the months of the year for the central Arctic (Fig. 27) and also for the area covered by the Norwegian, Barents, and Kara Seas. This would provide a check against heat-budget estimates and a basis for further study, of the interaction of the heat budget and atmospheric circulation. It would also be desirable to continue such calculations to cover several years in order to get an idea of year-to-year variations and their causes. As noted in Section VI there is reason to suspect that year-to-year differences in storm tracks and in amount of heat advection are substantial. One hope lies with the data reduced by the General Circulation Research Group at the Massachusetts Institute of Technology, under the direction of Professor Victor Starr. Data for about 700 radiosonde stations have been reduced for the years 1958--1962 inclusive to provide both meridional and zonal values of latent- and sensible-heat transport by the atmosphere. From these data it would be a relatively straightforward matter to compute transport across closed boundaries as outlined above. If year-to-year variations in heat budget can be estimated, based on data from drifting stations and aircraft reconnaissance, the comparison with variations in atmospheric circulation would be extremely interesting.

4.   Compilation of Mean Cloud Conditions Over the Arctic Basin

Routine weather-reconnaissance flights over the Arctic Basin have been conducted by the United States Air Force since 1947. Routine commercial airline flights have been conducted since 1956 (S.A.S.). Ice-reconnaissance flights have been conducted by the

U. S. Navy during much of this period. Yet when Vowinckel undertook
to compile maps of cloud conditions, he found it necessary to base
his maps mostly on <u>ground</u> observations from <u>land-based</u> weather stations.
tions.[7] Such data are very poor for this purpose because not only
are observations of middle and high clouds difficult from the ground,
but also conditions at land stations are not likely to be representa-
tive of the pack-ice area.

Cloud conditions influence both the longwave and the solar radia-
tion components critically, as explained in Section II. A systematic
effort should be made to recover as much data as possible from past
reconnaissance operations* and to review present procedures for data
collection and storage, with a view to recording both mean conditions
and year-to-year variations in a form readily useable for heat-budget
calculations.

5. Dynamics of Ice and Water Movement in the Arctic Basin

The heat exchange between the Arctic Ocean and other oceans
(Section VII) is about half due to ice export** and half to water
exchange. Both are subject to large year-to-year variations.

Estimates of the quantity of ice exported are based on estimates
of the speed and width of the East Greenland Current and of the mean
thickness of the ice being exported. The author believes that the
estimates of mean ice thickness constitute the greatest single
uncertainty in the heat budget of the Arctic <u>Ocean</u>, being consistently
underestimated by almost all investigators. With the advent of under-
ice submarine observations it should be possible to reduce this
uncertainty greatly. Preliminary reports[77][78] indicate that the
true mean thickness might be as much as 100 per cent greater than the
usual estimates.

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*A check with the meteorological data center at Ashville, North
Carolina revealed that cloud data from the "Ptarmigan" reconnaissance
flights is available for the period 1947 to 1965. Since 1952 these
flights have been scheduled <u>daily</u> over the Arctic Basin. Very little
use has been made of these data.

**When eventually melted, the ice represents a heat sink of 70--
80 cal/gm. Ice carried out by ocean currents can be thought of as the
export of heat deficit.

The amount of ice exported seems to vary greatly from year to
year; it is closely related to the mean wind patterns over the Arctic
Ocean. The rate of formation of ice is also closely related to the
winds, especially to strong and variable winds that break up the ice
and create open leads. The study of the relations between wind fields
and ice fields has already received a great deal of effort, but field
data relating convergence and divergence patterns of ice fields and
wind fields are very sparse and the relations are poorly understood.

With regard to water exchange, the greatest uncertainty concerns
the passage between Greenland and Spitsbergen, where normal oceano-
graphic sections are prevented by surface ice. How best to overcome
this difficulty is not clear; possibly by submerged buoys put in place
by submarine. This venture involves difficult operational and logis-
tic problems as well as unique political problems (Spitsbergen is
demilitarized by treaty, administered by Norway and largely occupied
by the Soviet Union). This problem is especially suitable for joint
consideration under international auspices.

Water exchange between Greenland and Iceland is also very uncer-
tain because of ice conditions; its study should be considered in
conjunction with that of the Spitsbergen section.

Between Iceland and the Shetland Islands numerous oceanographic
sections have been taken, and this work will no doubt be expanded
during the next few years. The bottom topography here is very
complicated, and the flow seems to be erratic; so there is much to be
done.

Flow through Bering Strait has been recorded electrically during
recent years by the Navy Electronics Laboratory. Similar methods
could be used for recording flow among the Canadian islands.

In summary, flow through all of the passages is rather poorly
recorded, but the greatest uncertainty is ice export past northeastern
Greenland. In the areas of water exchange, the greatest uncertainty
also is the flow past northeast Greenland.

6. Reducing Uncertainties in Longwave Components

Longwave Radiation Balance in Winter. Comparison of the radiation budget of the surface reported by various authors (Fig. 18) shows that the main discrepancy is in the net outgoing longwave radiation at the surface during the coldest months. This is mainly due to uncertainty about the intensity of back radiation from the atmosphere. As noted in Section II, the presence of clouds or other particles in the air during this season generally reduces the loss at the surface and may increase or decrease the loss to space.

A slight atmospheric turbidity due to ice crystals has often been observed during the coldest months. This turbidity does not constitute cloudiness, and does not greatly obstruct vertical vision (as with ice fog, horizontal visibility may be another matter).

Investigation of the vertical distribution of longwave radiation during winter by direct measurement would be desirable (rather than by computation from radiosonde data as by Marshunova). Such a study is prerequisite to optimum use of satellite observations.

Radiating Properties of Arctic Clouds. The longwave radiation components dominate the heat budget and are largely affected by the radiating properties of Arctic clouds. Marshunova reports that, at $-26°C$, Arctic clouds radiate only 85 per cent as a blackbody. [3] Vowinckel and Orvig [10] reject this report and assume blackbody radiation. It does appear that actual field investigation has been inadequate to support either view fully, but the difference is very important to the heat budget.

Areal Variation of the Temperature of the Surface. The measurement of the actual radiating temperature of the surface has always been difficult. For snow, it has been traditional practice to bury the thermometer bulb 1/2 below the surface. Obviously the presence of the thermometer disturbs the natural conditions, where gradients in in the top millimeter may be important. Moreover, during periods of maximum cold, steep vertical temperature gradients may exist in the lowest meter of the atmosphere. Since pressure ridges in the ice occupy a significant area and may rise to several meters, their

radiating characteristics could be significantly different from the level snow surface.

In summer, a great variety of shallow ponds may exist in the ice, with sharp thermal gradients near their surfaces. This greatly complicates mean conditions for latent- and sensible-heat exchange as well as radiation computations.

In the past these difficulties seemed insurmountable, but the rapid development of radiation-sensing instruments in recent years should greatly alleviate these problems[75][76] by making it possible to sample representative radiation temperature "profiles."

7. Evaporation and Turbulent-Heat Exchange

The determination of mean values for latent- and turbulent-heat fluxes is plagued by great instrumental and operational difficulties. It may be that much better direct observations of these quantities can be obtained by using radiation sensors. Regardless, better observation of variations in surface temperature over the pack will improve computations of mean values.

At the same time, independent determination of latent- and sensible-heat transport by the atmosphere (item 3), combined with a more reliable radiation balance for the earth--atmosphere system should give latent- and sensible-heat flux as a residual to a fair degree of accuracy.

Inspection of Table 10 (Section V) illustrates the enormous rate of heat loss from open leads in winter. All the values in Table 10, however, are computed from meteorological variables. Direct measurement of heat flux at the surface has never been feasible, especially over open leads. However, the contemporary development of photon-counting sensors offers hope of making such direct measurements. McAllister and Ewing (Scripps Institute) report measurements of the temperature gradient in the surface film of water -- measurements based on comparison of longwave emission at different wavelengths (and thus originating at different depths in the surface).

8. Interaction of Surface Heating Patterns and Cyclone Tracks.

As is discussed in Section VI, cyclone tracks in the North At
Atlantic are influenced by the pattern of surface-heat exchange. One
factor that influences this pattern is the position of the pack-ice
margin. Another is the pattern of ocean-surface temperature in the
North Atlantic and in the Norwegian and Barents Seas. Significant
anomalies in surface temperature of the ocean are common, and it would
be revealing to correlate the changing pattern of surface temperatures
with the movement of cyclones.

In the past, both factors were difficult to observe in detail,
for synoptic observations from these sea areas were sparse. In the
future, however, satellite sensors should provide excellent data for
studying this interaction. The HRIR (High Resolution Infrared) sen-
sors flown on the polar-orbit Nimbus I in 1964 provides an excellent
system for observing the progress of cyclones over ocean areas either
day or night. Satellite measurements of surface temperature, however,
involve substantial difficulties. Nimbus C will include a measurement
in the "window" region (10--11μ), which of course, will be obscured
by cloud and complicated by sea-state and boundary-layer problems.
In time, however, microwave radiometers will be included in satellite
instrumentation to provide surface temperature unobscured by cloud
cover.

In the meantime, extensive aircraft reconnaissance is being
flown over the North Atlantic and the Norwegian Sea. It would be
useful to review the possible basis for sea-temperature maps from this
data source. When satellite measurements do become available such
complementary data would facilitate interpretation and "calibration"
of the satellite data, thus reducing uncertainties concerning surface
emissivity, sea state, distribution of moisture and temperature near
the surface.

9. Paleoclimatic Observations in the Arctic

Our earth has gone through wide variations in climate during the
last few tens of thousands of years. One of the aims of the climatic
experiments described in this study is to reveal the basic causes and
interactions that were involved in these changes.

As yet, however, we have only rather crude ideas of the general climatic conditions associated with glacial ages. Atlantic Ocean bottom temperatures were apparently cooler during periods of ice accumulation, but no such change is apparent for the Pacific. Several other indicators suggest that a different pattern of water exchange existed between the Atlantic and Arctic Oceans, but the general picture of the climatic sequence at high latitudes is still obscure.

Whatever the true explanations may be, mountainous areas of Ellesmere Island and Axel Heiberg Island, which today still are largely covered by ice sheets, must have undergone wide changes in snow accumulation. Some of the reasons for believing this are given in Section VI. Conspicuous physical evidence includes the elaborate system of deep and long ice-carved fjords (found also in Norway, British Columbia, and Chile).

It should be possible to determine whether an open Arctic Ocean corresponded in time with the growth of ice sheets and whether accumulation ceased when the sea surface froze again. The glacial and land features of Spitsbergen have already yielded much evidence relevant to this question and similar investigation along the north side of Ellesmere Island can be expected to yield much more.

Techniques for dating ocean-bottom cores have been highly developed in recent years, and one would expect that paleoclimatic chronology would be more advanced than it really is.

In practice, however, it is difficult to obtain representative cores from the Arctic Basin because surface stations cannot maneuver but must move with the ice.

But Arctic research (compared to Antarctic research or geophysical research in general) has attracted support at only a low level. As a result field investigations have been limited severely, and this is the main difficulty.

10. Measures for Influencing the Arctic Heat Budget

Measures that might influence the Arctic heat budget must eventually be systematically investigated. It would be better to start now, at a modest level, than to be forced into hasty experimentation later. For one thing, the annual cycle permits field experiments in general, to be undertaken only once a year.

Among the measures that might be investigated to obtain quantita-
tive estimates of feasibility and effectiveness of modifying the heat
budget, the most obvious are measures to influence cloud cover and
surface albedo.

The ubiquitous stratus of summer is known to be almost ideally
suited for standard techniques of stratus dissipation by seeding. A
few crude tests have confirmed qualitative effectiveness, but systema-
tic tests have not been conducted to determine such basics as how
expensive it would be to achieve widespread effects or the best
methods for doing so. The possibility of inducing high cloudiness
also deserves attention because of its potential for affecting the
heat budget. This problem has apparently not been studied at all.
Influencing cloud conditions offers the potential for influencing the
heat budget in either direction.

With regard to altering surface albedo, there are at least
three lines of investigation that should be evaluated.

(a) One is the comparative effectiveness of various dark or
light substances on the surface (ash, soot, dirt, etc.). How does
a given substance behave? Does it collect into balls and melt down
into the surface, thus losing effectiveness? Can this be avoided?
What is the relative effectiveness of light and heavy layers? (The
systematic study of such questions as these would, at a minimum,
provide a better basis for evaluating the effects of such natural
events as volcanic eruptions. Even if the logistics of such measures
limit feasibility for large areas, systematic quantitative information
would be useful for various small-scale applications such as clearing
harbors and channels or preserving such ice structures as runways.)

(b) A second line of investigation is the evaluation of the use
of microorganisms for affecting albedo. It may be that seeding
microorganisms over a large area would be more feasible than spreading
other materials. This avenue has not been investigated.

(c). A third line of investigation would evaluate the cumulative
effects of momentary inundation of the ice (and snow cover) at various
times of the year. Such inundation could result from wave action

induced by explosives. Since producing explosions is one of the things we already know how to do, we could proceed directly to evaluation of the possible effects to be achieved by such methods.

In this connection it should be noted that the extreme density stratification of the upper layers of the Arctic Ocean (Fig. 40) provides an extraordinarily effective situation for the formation and transmission of internal waves at 300--400 meters depth. This phenomenon has been observed but has been little studied. Among the problems to be investigated are: the propagation characteristics of such waves, the effects on the mobility of pack ice (by causing fractures), the natural causes of internal waves, and the artificial generation of internal waves.

Appendix

MEAN MULTIYEAR MONTHLY RADIATION BALANCE

(Adapted from Marshunova)

MEAN MULTIYEAR MONTHLY RADIATION BALANCE

(in kcal/cm²)

Pack Ice Area

Grid Point	Radiation Components	I	II	III	IV	V	VI	VII	VIII	IX	X	XI	XII	Net Annual
90°N	Total solar	...	...	...	8.1	17.6	19.2	13.5	8.3	2.2	...	...	...	68.9
	Reflected solar	...	...	...	6.5	14.4	15.0	8.6	5.7	1.8	...	...	...	52.0
	Absorbed solar	...	...	...	1.6	3.2	4.2	4.9	2.6	0.4	...	...	...	16.9
	Net outgoing long-wave	-2.0	-1.9	-1.9	-2.2	-1.5	-1.2	-1.1	-1.1	-1.2	-1.3	-1.9	-2.1	-19.4
	Balance	-2.0	-1.9	-1.9	-0.6	1.7	3.0	3.8	1.5	-0.8	-1.3	-1.9	-2.1	-2.5
85°N 0°E	Total solar	...	...	0.6	8.5	17.6	19.2	13.5	8.4	2.5	...	...	...	70.3
	Reflected solar	...	...	0.5	6.9	14.4	15.2	8.6	5.8	2.1	...	...	...	53.5
	Absorbed solar	...	...	0.1	1.6	3.2	4.0	4.9	2.6	0.4	...	...	...	16.8
	Net outgoing long-wave	-2.0	-1.9	-2.0	-2.1	-1.4	-1.2	-0.9	-1.1	-1.0	-1.3	-1.9	-2.0	-18.8
	Balance	-2.0	-1.9	-1.9	-0.5	1.8	2.8	4.0	1.5	-0.6	-1.3	-1.9	-2.0	-2.0
85°N 60°E	Total solar	...	...	0.6	8.4	17.6	19.2	13.5	8.2	2.5	...	...	...	70.0
	Reflected solar	...	...	0.5	6.8	14.4	15.0	8.6	5.6	2.1	...	...	...	53.0
	Absorbed solar	...	...	0.1	1.6	3.2	4.2	4.9	2.6	0.4	...	...	...	17.0
	Net outgoing long-wave	-2.0	-1.9	-2.1	-2.1	-1.3	-1.2	-0.9	-1.0	-1.0	-1.3	-1.8	-2.1	-18.7
	Balance	-2.0	-1.9	-2.0	-0.5	1.9	3.0	4.0	1.6	-0.6	-1.3	-1.8	-2.1	-1.7
85°N 120°E	Total solar	...	...	0.6	8.4	17.6	19.2	13.5	8.4	2.6	...	...	...	70.3
	Reflected solar	...	...	0.5	6.8	14.4	15.0	8.6	5.8	2.2	...	...	...	53.3
	Absorbed solar	...	...	0.1	1.6	3.2	4.2	4.9	2.6	0.4	...	...	...	17.0
	Net outgoing long-wave	-2.3	-2.1	-2.1	-2.4	-1.5	-1.0	-0.9	-0.9	-0.9	-1.3	-1.8	-2.4	-19.6
	Balance	-2.3	-2.1	-2.0	-0.8	1.7	3.2	4.0	1.7	-0.5	-1.3	-1.8	-2.4	-2.6

(Table continued.)

Month

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Grid Point	Radiation Components	Month												Net Annual
		I	II	III	IV	V	VI	VII	VIII	IX	X	XI	XII	
85°N 180°E	Total solar	...	...	0.6	8.6	17.6	19.2	13.5	8.4	2.6	...	...	...	70.5
	Reflected solar	...	...	0.5	7.0	14.4	15.0	8.6	5.8	2.2	...	...	...	53.5
	Absorbed solar	...	...	0.1	1.6	3.2	4.2	4.9	2.6	0.4	...	...	...	17.0
	Net outgoing long-wave	-2.0	-1.9	-2.0	-2.4	-1.6	-1.2	-0.9	-0.9	-1.0	-1.4	-2.1	-2.2	-19.6
	Balance	-2.0	-1.9	-1.9	-0.8	1.6	3.0	4.0	1.7	-0.6	-1.4	-2.1	-2.2	-2.6
85°N 120°W	Total solar	...	...	0.6	8.6	17.8	19.2	14.0	8.5	2.6	...	...	...	71.3
	Reflected solar	...	...	0.5	7.0	14.6	15.0	9.0	5.9	2.2	...	...	...	54.2
	Absorbed solar	...	...	0.1	1.6	3.2	4.2	5.0	2.6	0.4	...	...	...	17.1
	Net outgoing long-wave	-2.1	-1.9	-2.0	-2.3	-1.7	-1.2	-1.2	-1.1	-1.1	-1.5	-2.0	-2.2	-20.3
	Balance	-2.1	-1.9	-1.9	-0.7	1.5	3.0	3.8	1.5	-0.7	-1.5	-2.0	-2.2	-3.2
80°N 135°E	Total solar	...	...	1.8	10.1	17.5	19.1	13.7	8.6	3.6	0.4	...	...	74.8
	Reflected solar	...	...	1.5	8.2	14.4	15.0	8.8	5.9	3.0	0.3	...	...	57.1
	Absorbed solar	...	...	0.3	1.9	3.1	4.1	4.9	2.7	0.6	0.1	...	...	17.7
	Net outgoing long-wave	-2.3	-2.3	-2.5	-2.5	-1.5	-1.0	-1.1	-0.9	-0.8	-1.3	-2.0	-2.7	-20.9
	Balance	-2.3	-2.3	-2.2	-0.6	1.6	3.1	3.8	1.8	-0.2	-1.2	-2.0	-2.7	-3.2
80°N 165°E	Total solar	...	...	1.9	10.0	17.7	19.0	14.0	8.6	3.6	0.4	...	...	75.2
	Reflected solar	...	...	1.5	8.1	14.5	15.0	9.0	5.9	3.0	0.3	...	...	57.3
	Absorbed solar	...	...	0.3	1.9	3.2	4.1	5.0	2.7	0.6	0.1	...	...	17.9
	Net outgoing long-wave	-2.4	-2.2	-2.5	-2.4	-1.6	-1.0	-1.2	-0.8	-1.1	-1.4	-2.0	-2.5	-21.1
	Balance	-2.4	-2.2	-2.2	-0.5	1.6	3.1	3.8	1.9	-0.5	-1.3	-2.0	-2.5	-3.2
80°N 165°W	Total solar	...	...	1.9	9.9	17.7	19.0	14.0	8.6	3.7	0.4	...	...	75.2
	Reflected solar	...	...	1.6	8.0	14.4	14.8	9.0	5.9	3.1	0.3	...	...	57.1
	Absorbed solar	...	...	0.3	1.9	3.3	4.2	5.0	2.7	0.6	0.1	...	...	16.1
	Net outgoing long-wave	-2.3	-2.2	-2.3	-2.2	-1.6	-1.0	-1.3	-0.8	-1.1	-1.4	-1.9	-2.2	-20.3
	Balance	-2.3	-2.2	-2.0	-0.3	1.7	3.2	3.7	1.9	-0.5	-1.3	-1.9	-2.2	-2.2

(Table continued.)

Grid Point	Radiation Components	Month												Net Annual
		I	II	III	IV	V	VI	VII	VIII	IX	X	XI	XII	
80°N 135°W	Total solar	...	...	1.9	9.9	17.6	19.2	14.2	9.0	3.7	0.4	...	...	75.9
	Reflected solar	...	...	1.6	8.0	14.4	15.0	9.1	6.2	3.1	0.3	...	...	57.7
	Absorbed solar	...	...	0.3	1.9	3.2	4.2	5.1	2.8	0.6	0.1	...	...	18.2
	Net outgoing long-wave	-2.2	-2.0	-2.3	-2.1	-1.6	-1.2	-1.1	-1.0	-1.2	-1.4	-2.1	-2.1	-20.3
	Balance	-2.2	-2.0	-2.0	-0.2	1.6	3.0	4.0	1.8	-0.6	-1.3	-2.1	-2.1	-2.1
75°N 160°E	Total solar	...	0.4	3.6	10.0	16.6	18.5	13.7	9.2	5.2	1.1	...	...	78.3
	Reflected solar	...	0.3	3.0	8.1	13.6	14.4	8.7	6.3	4.4	0.9	...	...	59.7
	Absorbed solar	...	0.1	0.6	1.9	3.0	4.1	5.0	2.9	0.8	0.2	...	...	18.6
	Net outgoing long-wave	-2.5	-2.3	-2.5	-2.4	-1.5	-1.0	-1.1	-1.1	-0.9	-1.0	-2.2	-2.5	-21.0
	Balance	-2.5	-2.2	-1.9	-0.5	1.5	3.1	3.9	1.8	-0.1	-0.8	-2.2	-2.5	-2.4
75°N 180°E	Total solar	...	0.4	3.6	10.0	16.8	18.5	13.5	9.2	5.2	1.1	...	...	78.3
	Reflected solar	...	0.3	3.0	8.1	13.6	14.4	8.7	6.3	4.4	0.9	...	...	59.7
	Absorbed solar	...	0.1	0.6	1.9	3.0	4.1	4.8	2.9	0.8	0.2	...	...	18.6
	Net outgoing long-wave	-2.5	-2.3	-2.5	-2.4	-1.5	-1.0	-1.1	-1.1	-0.9	-1.0	-2.2	-2.5	-21.0
	Balance	-2.5	-2.2	-1.9	-0.5	1.5	3.1	3.7	1.8	-0.1	-0.8	-2.2	-2.5	-2.6
75°N 160°W	Total solar	...	0.4	3.6	10.3	16.8	18.5	13.7	9.2	5.2	1.1	...	...	78.8
	Reflected solar	...	0.3	3.0	8.3	13.8	14.4	8.7	6.3	4.4	0.9	...	...	60.1
	Absorbed solar	...	0.1	0.6	2.0	3.0	4.1	5.0	2.9	0.9	0.2	...	...	18.7
	Net outgoing long-wave	-2.3	-2.2	-2.4	-2.2	-1.6	-1.1	-1.2	-1.1	-1.0	-1.1	-2.0	-2.5	-20.7
	Balance	-2.3	-2.1	-1.8	-0.2	1.4	3.0	3.8	1.8	-0.2	-0.9	-2.0	-2.5	-2.0
75°N 140°W	Total solar	...	0.4	3.6	10.3	16.8	18.7	13.7	9.2	5.3	1.1	...	...	79.1
	Reflected solar	...	0.3	3.0	8.3	13.8	14.6	8.7	6.3	4.4	0.9	...	...	60.3
	Absorbed solar	...	0.1	0.6	2.0	3.0	4.1	5.0	2.9	0.9	0.2	...	...	18.8
	Net outgoing long-wave	-2.1	-2.2	-2.6	-2.2	-1.6	-1.5	-1.1	-1.1	-1.1	-1.3	-2.0	-2.4	-21.2
	Balance	-2.1	-2.1	-2.0	-2.0	1.4	2.6	3.9	1.8	-0.2	-1.1	-2.0	-2.4	-2.4

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